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ACTA MINERALOGICA-PETROGRAPHICA

Tomus XXXIII

SZEGED, HUNGARIA

1992

NOTE TO CONTRIBUTORS

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Methods, techniques, material studied, description of the area investigated, etc.
Results
Discussion or conclusions
Acknowledgement
Explanation of plates (if any)
Tables
Captions of figures (drawings, photomicrographs, etc.)

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The abstract cannot be longer than 500 words.

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The tables generally should not exceed the type-area of the journal, i.e. 12.5x18.5 cm. Foldouts can only exceptionally be accepted.

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SERIES NOSTRA AB INSTITUTIS MINERALOGICIS, GEOCHIMICIS PETROGRAPHICIS UNIVERSITATUM HUNGARICUM ADIUVATUR

Kiadja

a József Attila Tudományegyetem Ásványtani, Geokémiai és Kőzettani Tanszéke
H-6722 Szeged, Egyetem u. 2—6

Kiadványunk címének rövidítése
Acta Miner. Petr., Szeged

SOROZATUNK A MAGYARORSZÁGI EGYETEMEK ROKON TANSZÉKEINEK TÁMOGATÁSÁVAL JELENIK MEG

CONTENTS

Acta Mineralogica-Petrographica, Szeged, XXXIII, 5—6, 1992.

TO THE MEMORY OF PROF. DR. GY. GRASSELLY

1920—1991

Prof. Dr. GYULA GRASSELLY's dead on 13 November, 1991, was a serious loss not only for his family, friends and coo-workers, but for the Hungarian Academy of Sciences, for the University of Szeged, as well as for both of the Hungarian and international geology, too.

He was born on 4th July, 1920, in Szeged, where he was brought up and persued his studies. After having achieved a first-class school-leaving certificate, he matriculated at the Faculty of Chemistry and Natural History of the Univ. of Szeged and finished his studies successfully. He passed his last semester in Berlin, studied under Prof. RAMDOHR's direction and gained excellent experiences in the analysis of ores. In 1944 he became certificated teacher of chemistry and natural history.

From 1944 until 1989 he worked at the Department of Mineralogy, Geochemistry and Petrography of the Univ. of Szeged (later József Attila University) as assistant, first assistant, then assistant professor and in 1964 he was appointed to professorship. Between 1968—1986 he was the head of the Department.

Between 1968—1986 he was the editor-in-chief of *Acta Mineralogica-Petrographica*, while it developed to an international periodical. In Hungary even

to-day this is the only mineralogical-geochemical and petrological scientific review issued in English language.

When retiring after 46 years from the same place where he began his career his merits were acknowledged by the honouring title of "professor emeritus" on the part of the University, and by the "Flag of the People's Republic" on the part of the President of the Hungarian People's Republic.

In his own estimation his career was modified three times. First he worked in the field of mineralogy, dealing with the methodology of mineralogical and petrographic analysis. His results were sum up in his first book entitled "Methods of Mineral and Ore Analysis" and produced him the Ph.D. degree.

The second modification of his career was the so called "ionpotential" era. In 1959 he gained the "Academic Doctor Degree". The title of his dissertation was "The role and importance of complex anion potentials in geochemistry".

At the same time industrial request aroused his interest on the Hungarian manganese ore deposit. Realizing the numerous for-reaching and unsolved problems of this subject and recognizing the necessity of an international collaboration, in 1976 he established the Commission on Manganese in the frame of the Association on the Genesis of Ore Deposits. In 1974 he launched the project "Genesis of Manganese Ore Deposits" within the International Geological Correlation Programme (IGCP). As a consummation of his work he edited — with I. M. VARENTSOV's co-operation — a three-volume monograph titled "Geology and Geochemistry of Manganese: Mineralogy, Geochemistry, Methods" even today used and acknowledged by specialist of manganese ores throughout the world.

His third and last, but not less important research trend was the complex geochemical and mineralogical-petrographic analysis of sediments containing organic matter in order to study the role and the effect of different physical factors of their diagenese. By these investigations he defined markedly the Hungarian organic geochemistry.

In 1972 he was elected vice-president of the international Union for Geological Sciences (IUGS) at the XXIVth International Geological Congress, Montreal. In 1982 he was elected the director of IUGS Advisory Board for Research Development Programme and later a leading member of the ICL Bureau. He made good use of all these international scientific experiences and connections in the Advisory Board of the Institution of Cultural Connection, the Hungarian UNESCO Committee and in the Hungarian National Committee of PUGWASH and IUGS.

The Hungarian Academy of Sciences appreciated GY. GRASSELLY's scientific achievements and even in 1956 he became a member of the Geochemical Scientific Committee where later he held the presidency, too. Since 1976 he was an associated member and since 1982 an ordinary member of the Hungarian Academy of Sciences. In 1990 he was elected the president of the Dept. of Earth-Sciences of the Academy.

As a real resident of Szeged his positive local patriotism appeared in a number of scientific and social activities. He was first a member, then the president and later, till his death, the vice-president of the Regional Committee of the Academy of Sciences, Szeged. His last work for his town was the organization of the reconstruction of Szeged Centre of the Academy. Unfortunately, he could not live to see the inauguration of the renewed hall which treasures his memory in every corner.

Standing up from his seat, Prof. GRASSELLY went away as if he came back in some minutes to finish his interrupted work. As a matter of fact, he did not left anything unfinished, he lead a full life till the end.

Prof. T. SZEDERKÉNYI

Acta Mineralogica-Petrographica, Szeged, XXXIII, 7—11, 1992.

SOLID SOLUTION IN MINERAL NOMENCLATURE

E. H. NICKEL*

Vice-chairman, Commission of New Minerals and Mineral Names,
International Mineralogical Association

INTRODUCTION

Guidelines for mineral nomenclature recommended by the Commission on New Minerals and Mineral Names (CNMMN) of the International Mineralogical Association were summarized by NICKEL and MANDARINO (1987—1989) and published in most of the international mineralogical journals.

One aspect of mineral nomenclature that was not covered in the guidelines was the question of how members of solid-solution series should be named. This matter was initially discussed by the Nomenclature Subcommittee of the CNMMN, and the recommendations of that body were subsequently considered and modified by the full CNMMN membership. These deliberations culminated in a general consensus that is embodied in this paper. Although similar to brief recommendations published by the Commission on New Minerals and Minerals Names of the All-Union Mineralogical Society of the USSR Academy of Sciences (Zap.Vses. Min. Ob. **106** (1977), 686—687), it is considered appropriate to publish this paper because it covers the subject more comprehensively and because it has the approval of the CNMMN.

Mineralogists wishing to give names to members of known solid-solution series are advised to adhere to the recommendation in this paper. However, to avoid confusion, mineral names or definitions already in the literature that contravene the recommendations should not be changed unless there are compelling reasons to do so, and then only if approved by a formal vote of members of the CNMMN.

Although general guidelines are recommended, readers will note that a certain degree of flexibility is permitted in the case of partial solid-solution series. Proposals for mineral names in this category will be judged by members of the CNMMN on the merits of each particular case.

Solid solutions can be considered in terms of three categories: complete solid solutions without structural ordering, solid solutions with structural ordering, and partial solid-solutions. Mineral nomenclature in each of these categories is discussed below.

COMPLETE SOLID-SOLUTIONS WITHOUT STRUCTURAL ORDERING

For purposes of nomenclature, a complete solid-solution series without structural ordering of the ions defining the end members is arbitrarily divided at 50 mole %, and the two portions are given different names, with each name

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applying to the compositional range from the end member to the 50% mark. For the sake of brevity this will be called the "50% rule". In *Fig. 1*, one name applies

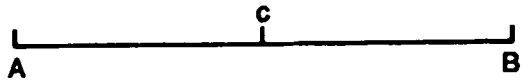

Fig. 1. Diagrammatic representation of a complete binary solid-solution series. A and B represent the two end members, and c represents the mid-point (50%).

to the range A-c, and the other to the range c-B. A mineralogical example of this is the forsterite-fayalite series $(Mg,Fe)_2SiO_4$, in which the name forsterite applied to the compositional range from Mg_2,SiO_4 to $MgFeSiO_4$, and the name fayalite applies from Fe_2SiO_4 to $MgFeSiO_4$.

Analogously, the 50% rule applied to members of ternary solid-solution series implies that mineral names should be given only to the three end members; each name should apply to the compositional range from the end member to the nearest right bisectors of the sides of the composition triangle, as shown in *Fig. 2.*

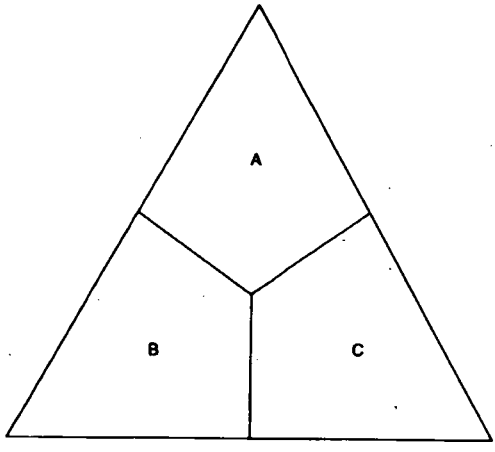

Fig. 2. Diagrammatic representation of a complete ternary solid-solution series. A, B and C represent the three compositional fields, each of which merits a mineral name.

For example, in the apatite series, $Ca_5(PO_4)_3(F,OH,Cl)$, the apices of the compositional triangle *(Fig.2)* can be represented by F, OH and Cl, respectively, making A = fluorapatite, B = hydroxylapatite, and C = chlorapatite.

According to the same principle, in a multi-component solid-solution series different mineral names can be given to isostructural or isotypic phases that have different chemical elements dominant in specified structural sites. An example of this is provided by minerals of the monazite series in which a number of different rare-earth elements can predominate in the cation structural site. The dominant element then specifies the appropriate "Levinson" suffix, e.g. monazite- (La).

SOLID SOLUTION WITH STRUCTURAL ORDERING

If there is structural ordering of the ions that define the end members within an otherwise disordered solid-solution series, the ordered phase may be given a mineral name different from those of the end members. An example is provided by dolomite, $CaMg(CO_3)_2$, in wich ordering of the Ca and Mg ions results in a crystal structure distinct from those of calcite and magnesite, the Ca and Mg end-members respectively, of the $(Ca,Mg)CO_3$ series. It is recommended that the name of a newly-described ordered phase discovered in an existing solid-solution series be derived from, or related to, the name of the solid or one of the end members, although the author of the name is not obligated to do so.

PARTIAL SOLID-SOLUTION SERIES

If there is limited solid solution at one or more of the end members, and the solid solution does not extend to the 50% mark, the 50% rule is generally applied. Therefore, in *Fig. 3*, the name of end-member A applies to the compositional range A-c and the name of end-member B applies to the range c-B, even if known compositions extend only to A' or B'; this is to allow for the possibility of new chemical data extending compositions toward c. For purposes of nomenclature, it does not matter whether or not A and B are isostructural.

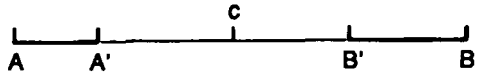

Fig. 3. Diagrammatic representation of a partial binary solid-solution series in which A'-B' represents the miscibility gap.

If the miscibility gap is to one side of the 50% mark, as in *Fig. 4*, and if the phases represented by A-A' and B-B' are not isostructural, a separate name should not be given for the range B'c if it is very small, but if it is of substantial extent, then a separate name might be justified. The dividing line between a "small" range and a "substantial" one, in this case and others given below, can be taken as about 10 mole %, although each situation should be regarded on its own merits.

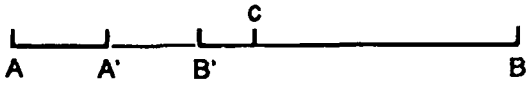

Fig.4. Diagrammatic representation of a partial binary solid-solution series in which A'-B' represents the miscibility gap, and the range B-B' encompasses the mid-point, c.

Similar considerations should apply to ternary or higher-order partial solid-solution series. Therefore, in a situation like that depicted in *Fig. 5*, the field defined by composition FGED does not warrant a separate name if it is very small, but may be given a separate name if it is of substantial size.

If the known compositions embrace the 50% mark, but do not appear to extend to either end member *(Fig. 6)*, only one name should apply to the compositional range. However, here again, the compositional range should be taken into account; if it is very small, then only one name should be given, but if it is large,

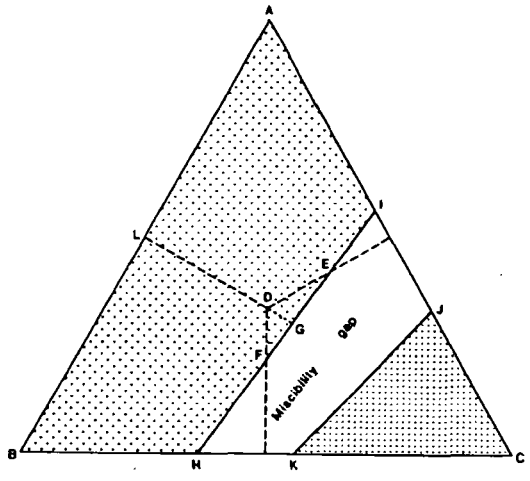

Fig. 5. Diagrammatic representation of a partial ternary solid-solution series in which the area HKJI represents the miscibility gap, and D represents the mid-point of the triangle.

consideration may given to two names. An example of a mineral in this category is pentlandite, $(Ni,Fe)_9S_8$, the composition of which centres around Ni:Fe = 1:1, and compositions near the Ni and Fe end-members are not known.

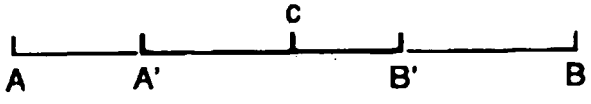

Fig. 6. Diagrammatic representation of a partial binary solid-solution series in which the solid solution is limited to the region A'-B'.

The analogous situation in a ternary solid solution can be represented by *Fig. 7*, in which known compositions cluster about a geometric boundary or boundaries. If the scatter of compositional points is small, only one name should be given to the cluster, but if the scatter is large, consideration can be given to more than one name.

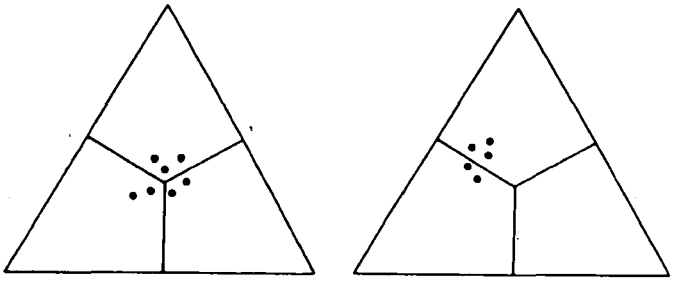

Fig. 7. Diagrammatic representation of ternary solid solution series in which known compositions cluster about geometric boundaries.

In cases such as those illustrated by *Figs. 6* and *7*, one particular composition of a type specimen should be nominated as the type, because later work might well reveal a wider range of compositions justifying two (or more) names. One of these should be the name already in existence.

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NICKEL, E. H. and MANDARINO, J. A.: Procedures involving the IMA Commission on New Minerals and Mineral Names, and guidelines on mineral nomenclature. Acta Petrologica et Mineralogica, **6**, 1987) 252—278 (in Chinese); American Mineralogist, **72**, (1987), 1031—1042; Boletin de la Sociedad Espanola de Mineralogia **12**, (1989), 1—30 (in Spanish); Bulletin de Mineralogie, **110**, (1987), 717—741; Canadian Mineralogist, **25**, (1987). 353—377; Fortschritte der Mineralogie, **65** (1987), 175—196; Indian Journal of Earth Sciences, **14**, (1986), 152—188; Mineralogical Journal, **13** (1987), 505—532; Mineralogical Magazine, **52** (1988), 275—292; Mineralogy and Petrology, **37** (1987), 157—179; Mineralogicheskiy Zhurnal, **11**, (1), (1989), 51—86 (in Russian); Rendiconti della Societa Italiana di Mineralogia e Petrologia, **4**, (1987), 27—53 (in Italian); Rivista Mineralogica Italiana, (1988), Supplemento al no. **1**, 5—31 (in Italian); Schweizerische Mineralogische und Petrographische Mitteilungen, **67**, (1987), 185—210.

Manuscript received, 14 October, 1991

Acta Mineralogica-Petrographica, Szeged, XXXIII, 13—24, 1992

DIAGENETIC NEOFORMATIONS IN MIDDLE TRIASSIC EVAPORITIC AND CARBONATE ROCKS, MECSEK MTS. (S. HUNGARY)

I. VICZIÁN*

Hungarian Geological Survey

ABSTRACT

In the Middle Triassic (Anisian) evaporitic and carbonate rocks of the Mecsek Mts. diagenetic neoformation of corrensite, Mg-rich chlorite, magnesite, albite and quartz has been observed by microscopic and X-ray diffraction methods.

Fine-grained magnesite is found in clay mineral aggregates of siltstones of detrital origin. Magnesites of very similar lithologic character have been found in the Drauzug area of the Eastern Alps and in Hesse, Germany, in a comparable stratigraphic position.

Authigenic albites in limestones occur as euhedral crystals oriented according to (010).

The formation of euhedral quartz of prismatic shape is accompanied by the formation of corrensite in carbonate rocks.

The stability relations of these minerals are discussed in terms of thermodynamic considerations of LIPPMANN. The possible effects of Mg-enrichment in the solution and of transformation reactions of detrital expandable clay minerals are discussed.

INTRODUCTION

The Middle Triassic (Anisian) sequence of the Mecsek Mts. was studied by using the rock samples of the borehole Pécs-IX (see *Fig. 1*). Clay minerals were described in a previous publication (VICZIÁN 1993). In the present paper the results of microscopic investigations are presented and genetic problems of the mineral neoformations are discussed.

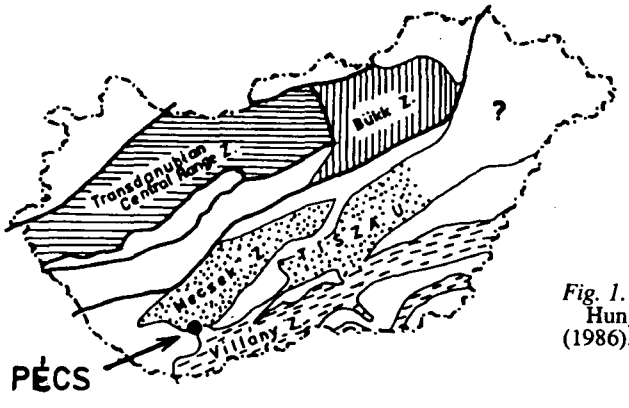

Fig. 1. Permian-Triassic facies units in Hungary according to HAAS *et al.* (1986). The location of the borehole at Pécs is shown.

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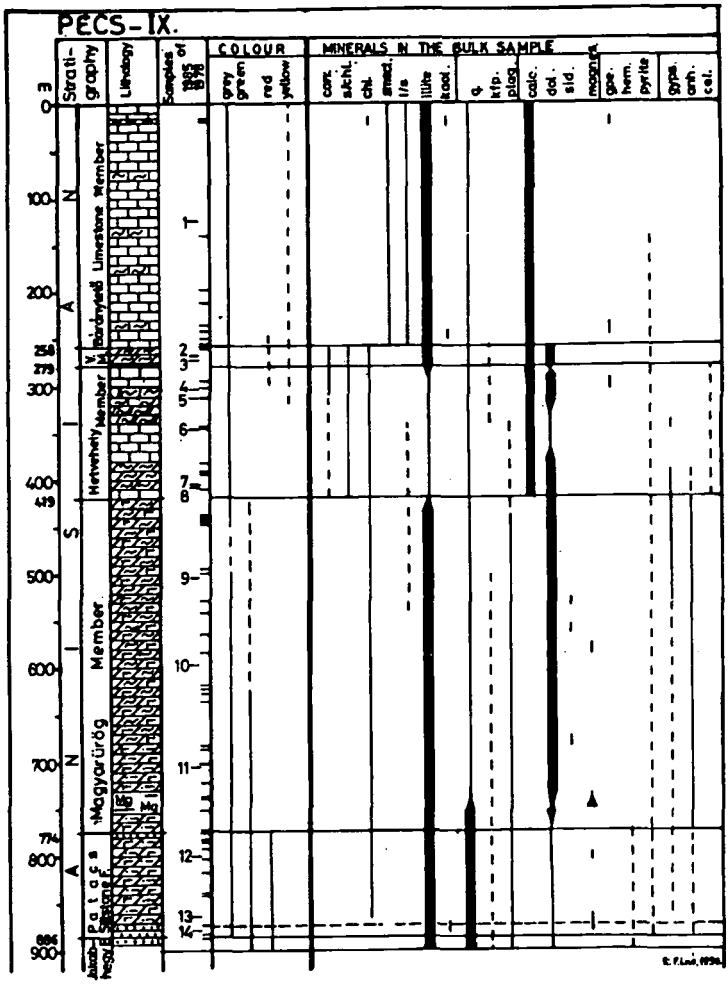
A microscopic study of the section was previously carried out by LELKES-FELVÁRI (1979, 1985). She observed textural features related to sedimentary environment and diagenetic processes. Detrital grains of acidic volcanics, shale, quartz, feldspar, biotite, chlorite and muscovite were found in the siltstone sequence. Neoformation of quartz and feldspar was described in nearly all rock types.

Lithologic column, stratigraphic subdivision and characteristic colour of the rocks are shown in *Fig. 2* on the basis of the results of previous investigations of


PÉCS-IX.
COLOUR
MINERALS IN THE BULK SAMPLE
Strati-graphy
Lithology
gray
green
red
yellow
corr.
s/chl.
chl.
smect.
i/s
illite
kaol.
q.
kfp.
plag.
calc.
dol.
sid.
magnes.
goe.
hem.
pyrite
gyps.
anh.
cel.
Member
Limestone
Member
Member
Member

LELKES-FELVÁRI (1979, 1983), RÁLISCH (1981) and ORAVECZ-SCHEFFER and GÓCZÁN (1986). A detailed geological description of the region was given by NAGY (1968). Bulk mineralogical compositon was determined by VICZIÁN (1978, 1985 , see *Fig.* 2) by X-ray diffraction analysis.

A brief macroscopic description of the rock samples investigated in the present study is given in Table 1.

TABLE 1.

Macroscopic description of the samples

1.	125.0—125.5 m	*Limestone*, dark grey to grey, hard.
2.	268.7—269.4 m	*Dolomitic marly limestone*, dark grey and light grey angular grains.
3.	272.5—274.0 m	*Calcitic dolomite*, light grey, brecciated, with light red, loose matrix.
4.	303.7—305.2 m	*Marly limestone*, dark grey with thin white limestone layers and lenses.
5.	312.4—312.6 m	*Marly limestone*, dark grey, hard, homogeneous.
6.	345.9—347.0 m	*Limestone*, dark grey, hard, calcitic veins.
7.	405.0—406.5 m	*Dolomite*, grey, hard, with thin, dark grey clay layers, calcite veins.
8.	408.0—410.4 m	*Limestone*, grey, hard, with thin, dark grey clay layers, calcite veins.
9.	503.0—505.0 m	*Anhydrite- and gypsum-bearing claystone*. Alternation of grey, dark grey clay and white, glittering sulphate layers.
10.	597.4—599.9 m	Alternation of thin a/ and b/ layers: a/ *Anhydrite-bearing dolomite marl*, dark grey with desiccation cracks. b/ *Gypsum- and dolomite-bearing siltstone*, grey, with micas and sporadic plant debris.
11.	704.8—707.3 m	*Dolomite marl*, grey, dark grey, layered, hard.
12.	799.4—800.4 m	*Siltstone*, greenish and reddish grey, dark grey, layered, with micas and brown magnesite (?) grains.
13.	864.2—864.8 m	*Anhydrite-bearing claystone*, greenish grey with pink fibrous anhydrite veins and magnesite (?) grains.
14.	878.0—878.6 m	*Siltstone*. Alternation of thin dark grey and light grey layers with slightly undulating surface.

METHODS

The investigations were carried out in the Institute of Mineralogy and Petrography of the Tübingen University, Germany, using laboratory methods of separation of grain fractions developed by LIPPMANN (see e.g. LIPPMANN and PANKAU, 1988).

Grains of the fraction 63 to 125 μm were scattered over a glass slide and mounted with canada balsam (n=1.54). Fractions containing high carbonate content (samples No. 2, 4, 9, 10/a and 10/b) were first examined in their original state and then treated with 0.5 n HCl. The insoluble residue of these fractions was mounted again on a glass slide with canada balsam.

Fractions were investigated under polarizing microscope and in special cases by X-ray diffraction (determination of albite in samples No. 6 and 8 and of magnesite in samples No. 12 and 13).

RESULTS

The results of the microscopic investigations are summarized in Tables 2 and 3.

The following types of mineral grains were found: aggregates of carbonate minerals including magnesite-bearing clay aggregates, carbonate-free clay aggregates, coarse-grained quartz + feldspar + mica + opaque aggregates as well as mineral grains of feldspars, quartz, micas (muscovite, biotite), pyrite, sparry carbonates, celestite, anhydrite and fluorite. In what follows, only aggregates of clay-carbonate composition, feldspar and quartz will be described in detail.

TABLE 2.

Type of carbonate grains, fraction 63 to 125 μm, untreated material, borehole Pécs-IX.

Stratigraphy	Sample No.	Fine-grained aggregates			sparry carbonates	carbonate minerals according to X-ray analysis
		carb.>clay (colour: light)	carb.>clay (colour: dark)	clay>carb.		
Bárányteto M.	1.	+++				dol.?
Hetvehely +Viganvár	2.		+++		+	dol.>calc.
	3.	+++			+	dol.
	4.		+++		+	calc.>dol.
Magyar-ürög M.	9.		+		+	magn., dol.?
	10/a.		+++			dol.
	10/b.		+		(+)	dol.
Patacs F.	12.			+++		magn.
	13.			+	+	magn.
	14.				(+)	—

+++	high	calc.:	calcite
+	medium	dol.:	dolomite
+	low	magn.:	magnesite
(+)	very low		

Carbonate - clay aggregates including magnesite-bearing clay aggregates

Aggregates containing mainly carbonate minerals could be observed in samples where carbonate minerals were not dissolved by HCl (Table 2). The aggregates are sometimes build up of fine micritic carbonate grains sometimes resembling rhombohedrons. The carbonate aggregates contain clay minerals which are set free

when treated with HCl. The proportion of carbonate to non-carbonate material within the aggregates varies in the section. In the aggregates of samples No. 12 and 13, carbonate (magnesite) is less abundant than non-carbonate clay material. Aggregates of sample No. 14 are carbonate-free even in the untreated material. In other samples carbonate aggregates contain only little non-carbonate material.

TABLE 3.

Results of the microscopic investigation of the fraction 63 to 125μm, non carbonate minerals (carbonate-free samples or samples treated with 0.5n HCl), bore hole Pécs-IX

Stratigrahpy	Sample No.	Fine-grained aggregates (dark grey, clay±pyrite)	Coarse-grained aggr. (q+fp+mica+opaque)	Quartz (detrital)	Quartz (euhedral)	Quartz (thin platelets)	Feldspar (detrital)	Feldspar (authigenic albite)	M ica (Muscovite)	Mica (biotite-Chlorite)	Pyrite	Celestite
Bárányteto M.	1.		(+)			(+)			(+)			
Hetvehely+ Vigan-vár Members	2.	+	+		+		+			(+)	+	
	3.		+	+	(+)	+	(+)			(+)	+	(+)
	4.			+	+++		+				+	
	5.	+		+			(+)	+			+	
	6.	+					+	+			(+)	
	7.	+			(+)	(+)	+	(+)	(+)			+
	8.	+			(+)			+	(+)		+	
Magyarü-rög M.	9.	+	+	+	+		+					
	10/a.	+++	+	+			+				+	
	10/b.	+	+++	+		(+)	+		+			
	11.	+++			(+)		(+)		+	(+)		
Patacs F.	12.	+++					+		(+)			
	13.	+	+	+	+		+		(+)	(+)		(+)
	14.	+++		+					(+)	(+)		

+++ high
+ medium
+ low
(+) very low

q: quartz
fp: feldspar

As determined by comparison with X-ray analysis of the bulk samples most carbonate grains are dolomite. In samples No. 9, 12 and 13 magnesite was found by X-ray analysis. According to the shift of the reflection 104 toward higher d

values approximately one fourth to one third of Mg^{2+} is replaced by other ions, most probably Fe^{2+}:

	2Θ	d
magnesite, JCPDS card No. 8-479:	32.65°	2.742 Å
samples No. 12 and 13:	32.5 °	2.76 Å
siderite, JCPDS card No. 29-696:	32.02°	2.795 Å.

A similar shift of the reflections 113 and 116 can be observed.

Magnesite has been found in the Eastern Alps and in Germany in rocks having similar stratigraphic and environmental relations. In particular, in the Drauzug area of the Eastern Alps magnesite is a normal constituent of terrigeneous and evaporitic sedimentary rocks at the boundary of Upper Scythian - Lower Anisian in a zone extending 140 km (NIEDERMAYR *et al.* 1980, NIEDERMAYR 1989, NIEDERMAYR. *et al.* 1989). Magnesites found here may have Fe contents up to 20 weight per cent. Their genesis is related to deposition in "a broad coastal plain under transitional conditions between a fluviatile setting and a coastal sabkha environment" (NIEDERMAYR *et al.* 1989, p. 55). In Germany, magnesite was determined by X-ray diffraction in gypsum-bearing silty sediments in a profile at Borken, Hesse (BÜHMANN and RAMBOW 1979). The age of these rocks is Upper Spathian (Upper Buntsandstein, Rötsalinar) close to the boundary of Anisian (DOUBINGER and BÜHMANN, 1981).

Feldspars

Feldspars occur both as detrital grains and euhedral crystals. Feldspars were identified by their interference figure, optic sign, extinction angle and variation of index of refraction in relation to canada balsam.

Few grains of *detrital feldspar* of irregular shape and inhomogeneous extinction can be observed in nearly all samples. Both potassium feldspar and plagioclase occur, twinning is very rare. The grains are cloudy, altered and contain inclusions of carbonates, sometimes also pyrite and micas.

In limestone samples No. 5, 6 and 8 many *euhedral feldspar* crystals with well-developed crystal faces can be observed (*Figs. 3* and *4*). The rounded internal core of these crystals is altered and contains inclusions of carbonate grains, "sericite", pyrite and yellowish brown or black micolites. The core is surrounded by a clear overgrowth of the same optical orientation. No difference in the extinction angle of the core and the rim could be observed. Euhedral feldspars are untwinned. Roc Tourné type twins as described by FÜCHTBAUER (1948) cannot be observed because almost all crystals are oriented parallel to (010). The form of the crystals is usually an elongated hexagon formed by the traces of crystal faces (001), (110) and (111). Less frequently grains in other orientations, probably with the face (001) parallel to the plane of the preparation, occur.

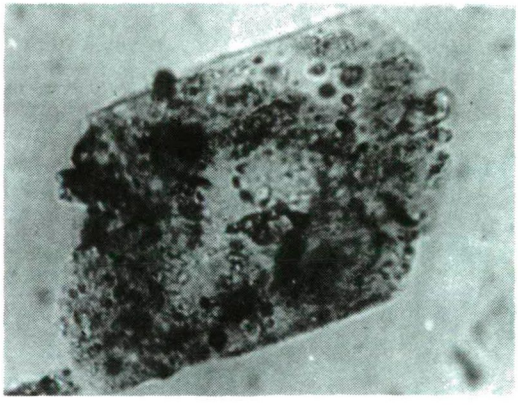
The index of refraction is lower than that of canada balsam. The relief is low but varies with the orientation of the grain if the crystal face (001) is nearly parallel to the plane of the preparation. Extinction is parallel to the elongation on (001) faces and oblique on (010) faces. Biaxial positive extinction figures with large axial angles can be observed on faces (010). Interference colours are most frequently yellow but vary from first order grey to blue. The optical properties can be interpreted as those of *low-albite* (ordered). As measured by extinction angles of the few albite twins observed the mineral may contain about 10 per cent anorthite molecule.

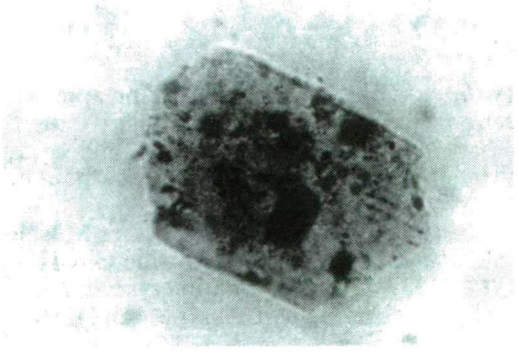
X-ray diffraction analysis was made of the sampes No. 6 and 8 in order to prove the optical determination. The grains of the fraction 63 to 125 μm of the material previously treated with 0.5 n HC1 were mounted on a glass slide with few drops of distilled water. The feldspar reflections are clearly those of low-albite (as compared with the JCPDS card No. 20-554). The intensities of the following reflection are clearly stronger than they should be in a powder free of preferred orientation:

d (Å)	hk1
6.38	001, 020
3.19	002, 040
2.125	060
1.594	080.

This is in agreement with the optical observation that most albite crystals have faces (010) or (001) parallel to the plane of preparation.

Authigenic albites were frequently reported from limestones and related rocks, e.g. from Liassic-Dogger of Zweisimmen, Switzerland (FÜCHTBAUER 1948), Muschelkalk of Göttingen (FÜCHTBAUER 1950), Zechstein near Braunschweig (FÜCHTBAUER 1956). Similar authigenic overgrowths of albite were observed in gypsum and gypsum-bearing claystone in the Muschelkalk of Nagold by LIPPMANN and PANKAU (1988).



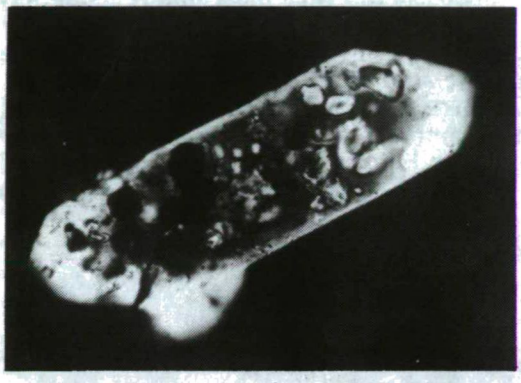


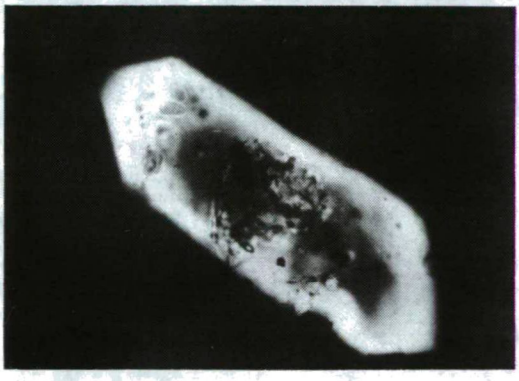
Figs. 3 and *4.* Authigenic albites from sample No. 2, dolomitic marly limestone, borehole Pécs-IX, 268.7 to 269.4 m, Anisian, Viganvár Member. || N, 500X (left), 300X (right).

Quartz

Few *detrital* quartz grains of irregular shape were observed in almost all samples except pure carbonate rocks.

Euhedral quartz of regular prismatic shape terminating on one or both ends by rhombohedrons is very frequent in the insoluble residue of the limestone samples No. 2 and 4 but it occurs in many other samples too. The external rim of these crystals is clear but the core contains many inclusions such as carbonates and brownish-black opaque microlites (*Figs. 5* and *6*). This variety of quartz is considered to be authigenic. In some cases very small prismatic overgrowths can be observed on the surface of clay mineral aggregates (sample No. 13) or overgrowths of irregular shape on detrital quartz grains (samples No. 3, 10/a and 13) may occur.





Figs. 5 and *6*. Authigenic quartzes from sample No. 6, limestone, borehole Pécs-IX, 345.9 to 347.0 m, Anisian, Hetvehely Member. +N, 300X.

Another variety of quartz are very *thin platelets* of irregular shape apparently perpendicular to the optic axis in direction c giving clear uniaxial positive interference figures. Also the plates may have inclusions of carbonate and opaque minerals. This variety is most abundant in sample No. 3 (so-called "boundary dolomite").

Euhedral prismatic quartz crystals were described from the Upper Muschelkalk at Haigerloch by LIPPMANN and SCHLENKER (1970) and from the Gipskeuper at Pfäffingen by LIPPMANN and STEINER (1983). Both prismatic crystals and thin plates were found in the Middle Muschelkalk at Nagold by LIPPMANN and PANKAU (1988).

Magnesite

The magnesites of the Drauzug area are considered by NIEDERMAYR (1989) to be converted from preexisting carbonates in an early stage of diagenesis. The genesis of magnesite will be considered here from the point of view of stability relations of rhombohedral carbonates in aqueous solutions (LIPPMANN 1980, 1982a, b).

Considering the stable phase diagram for the system $CaCO_3$ - $MgCO_3$ - H_2O at 25 °C (LIPPMANN, 1982a) magnesite should crystallize from solutions having activity fractions $X_{1\,[Mg2+]}$ higher than 0.832. Magnesites should crystallize even from normal sea water because in sea water $X_{1\,[Mg2+]} = 0.865$ (LIPPMANN 1982b). The formation of magnesite, however, is inhibited at ordinary temperature by the hydration of Mg^{2+} ions, even if seeds of magnesite are present in the solution. Even at elevated activity fractions of Mg^{2+} either aragonite or hydrous magnesium carbonates such as nesquehonite, $MgCO_3.3H_2O$, or hydromagnesite, $Mg_5(CO_3)_4(OH)_2.4H_2O$, precipitate firstly.

The transformation of these metastable phases into magnesite takes place first at elevated temperatures. Also huntite, $CaMg_3(CO_3)_4$, may be a metastable precursor of magnesite (LIPPMANN 1982a and LIPPMANN 1980).

As the activity fraction $X_{1[Mg2+]}$ of sea water, 0.865, is near to the eutectic point of the dolomite-magnesite pair, 0.832, a further enrichment of Mg^{2+} ions in the solution may favour the precipitation of magnesite. Simultaneous precipitation of gypsum may extract Ca^{2+} ions from the solution and produce high Mg^{2+} to Ca^{2+} ratios. This may be the case in the sequence studied where magnesite is accompanied with gypsum and anhydrite layers.

On the other hand, no concurrent formation of Mg-rich silicates such as chlorite, corrensite, talc, sepiolite or palygorskite could be observed in the portion of the sequence containing magnesite. All these minerals would be able to form as stable or metastable phases from high magnesian solutions according to the stability relations of these minerals in the system MgO - Al_2O_3 - SiO_2 - H_2O (LIPPMANN 1979). The crystallization of clinochlore (chlorite, corensite) may not have taken place due to insufficient Al_2O_3. No sepiolite or attapulgite were formed instead, probably because of low activity of silica. Talc forms very slowly at low temperatures.

The high iron content presumed to be present in the magnesites studied can be explained by the solubility relations in the system $MgCO_3$ - $FeCO_3$ - H_2O (LIPPMANN 1980 and LIPPMANN 1982b). By comparison of the solidus and solutus lines in the slubility diagrams of this system it is evident that high Fe concentrations in the solid are in equilibrium with rather low concentrations of Fe in the solution. In other words Fe readily precipitates from solution together with Mg to form carbonates. The high ferrous iron concentration is a consequence of the reducing conditions during diagenesis.

Albite

Albite was found in limestones of the Hetvehely Member deposited in sea water of normal salinity. Considering the stability relations in the system K_2O -Al_2O_3 - SiO_2 - H_2O (LIPPMANN 1982c) and in the system Na_2O -Al_2O_3 - SiO_2 - H_2O (unpublished digram of LIPPMANN see *Fig. 7*) it is obvious that silica must be enriched in the solution as compared with normal sea water in order to reach

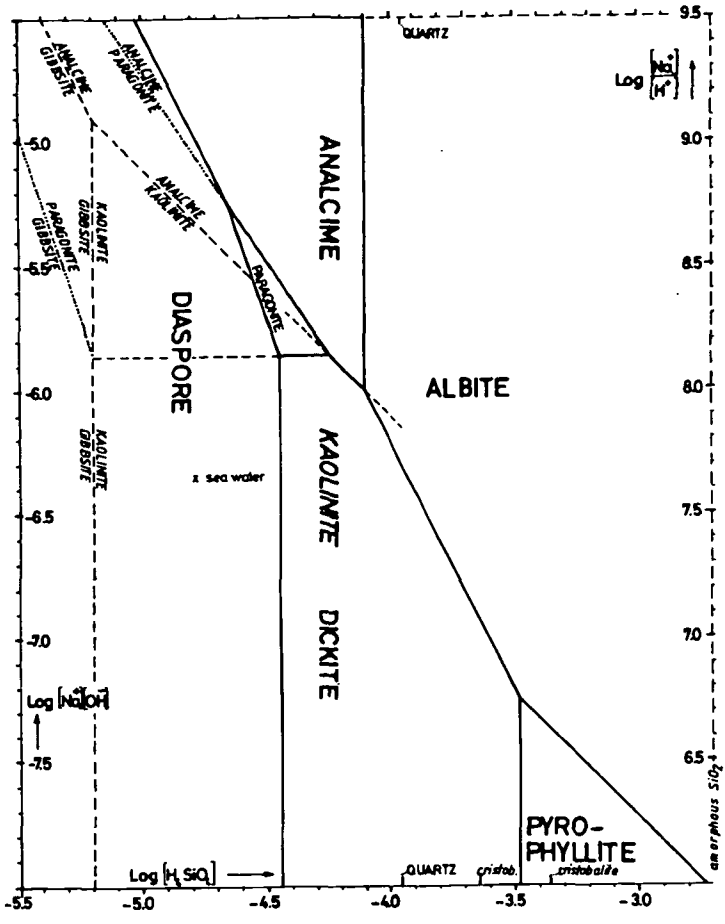


Fig. 7. The system Na_2O - Al_2O_3 - SiO_2 - H_2O at 25 °C with Al_2O_3 as immobile component calculated by LIPPMANN (unpublished diagram, improved version on *Fig. 4* in LIPPMANN 1979). Explanations: CAPITALS: stable minerals. *ITALICS:* metastable minerals reacting in metastable equilibrium; *italics*: unstable minerals; continuous lines: stable phase boundaries; dashed and dotted lines: metastable phase boundaries.

the stability fields of feldspars. In the presence of sufficient silica, the type of the feldspar crystallized depends on the K+ and Na+ activity in the solution. The point for the composition of sea water is near the boundary of the stability fields of plagioclase and K-felsdpar in the diagram $[K^+]$. $[OH^-]$ versus $[Na^+]$. $[OH^-]$ (unpublished diagram of LIPPMANN see *Fig. 8*). The composition of the solution may easily move into the stability field of albite if e.g., potassium is removed from solution by the reorganization of poorly crystalline micas or mixed-layer illite/smectites.

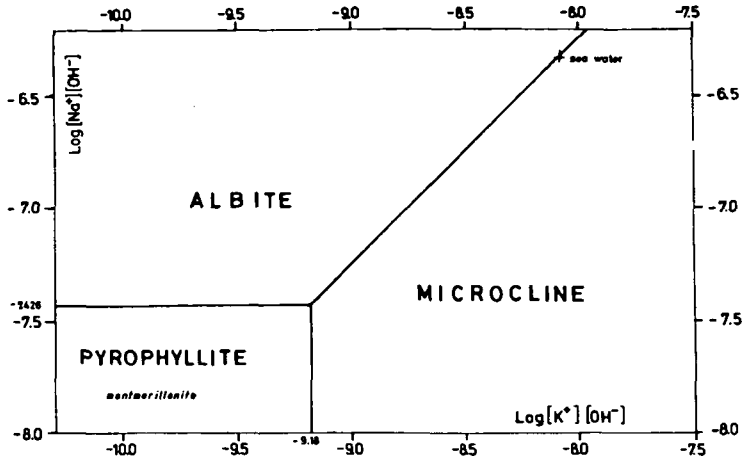


Fig. 8. The system K_2O - Na_2O - Al_2O_3 - SiO_2 - H_2O at 25 °C with Al_2O_3 as immobile component and log $[H_4SiO_4]$ = -3.3 (30 mg/l SiO_2).Unpublished diagram constructed by LIPPMANN (1988) using the thermodynamic data of ROBIE *et al.* (1978).
Explanations: CAPITALS: stable minerals, *italics*: unstable mineral. Note that the composition of sea water coincides with the boundary of the stability fields of albite and microcline within the accuracy limits of the data.

Quartz

Neofromation of minor amounts of quartz has been observed in nearly all samples but authigenic quartz crystals are frequent only in marly limestones of the Hetvehely and Viganvár Members. The neoformation of quartz coincides in these cases with the occurrence of corrensite in the clay fraction. It is probable that corrensite has formed by reaction of a smectite-like detrital clay material with solutions rich in Mg^{2+} and alkalinity (VICZIÁN 1993). The model of this transformation is the reaction of the end member pyrophyllite to clinochlore (Table 5 in LIPPMANN, 1979):

$$Al_2Si_4O_{10}(OH)_2 + 5\ Mg^{2+} + 10\ OH^- \rightleftarrows Mg_5Al_2Si_3O_{10}(OH)_8 + SiO_2 + 2\ H_2O.$$

pyrophyllite clinochlore quartz

Quartz is a product of this reaction. Another source of silica may be the transformation smectite to illite which can be described by the reaction of the end member pyrophyllite to muscovite (Table 2 in LIPPMANN 1979):

$$3\ Al_2Si_4O_{10}(OH)_2 + 2\ K^+ + 2\ OH^- + 10\ H_2O \rightleftarrows 2\ KAl_3Si_3O_{10}(OH)_2 + 6\ H_4SiO_4.$$

pyrophyllite muscovite

Thus both the formation of corrensite and illite may produce quartz during diagenesis.

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ZEOLITE MINERALS FROM INTERMEDIATE VOLCANIC ROCKS OF TOKAJ MTS., NE-HUNGARY

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Herman Ottó Museum

ABSTRACT

In recent years zeolite minerals have been found in large number in the Tokaj Mts. The examinations show that the parageneses in the intermediate rocks differ from those known in the acid rocks or pyroclastites of the area. The dominating minerals are mordenite, heulandite and stellerite. As a rarity two species the hungarian occurence of which was unknown up to now were found in an outcrop near Regéc. In this paper the author gives the description of the parageneses, the species identified, discusses the formation of the assemblages and crystal morphology referring to instrumental examinations.

INTRODUCTION

There are very few data presented on zeolites occurring in the intermediate volcanic rocks of the Tokaj Mts. As first TOKODY (1959) described heulandite and chabasite crystals found in the cavities of andesite in a quarry near Füzérkomlós. In the sixties G. PANTÓ collected heulandite crystals occurring in andesite, near Ujhuta (KULCSÁR oral communication). The exact place of the locality is not known however. In the course of the recently commenced geochemical research of the old workings at Telkibánya white chabasite rhombohedras reaching half centimetre were found in the Teréz-adit (FEGYVÁRI oral communication). JÁNOSSY *et al.* (1987) referred to mesolite and stilbite from the Mt. Kopasz, near Tokaj, without further data. SZAKÁLL (1991) wrote up a locality with abundant zeolites at Óhuta and Regéc.

METHODS

The equipment used for the examinations: SIEMENS D 500 (Dept. of Mineralogy, ELTE, Budapest) and PHILIPS PW 1730/10 (Dept. of Mineralogy, University of Veszprém) X-ray diffractometers. EDAX 9900 energy dispersive spectrometer and AMRAY 1830 I scanning electron microscope (Dept. of Metallurgy, University of Miskolc).

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LOCALITIES AND MINERALS

Óhuta

The richest zeolite occurrence is in an outcrop on Mt. Kis-Zabarla by a dirt track along the right side tributary of the Hutavölgyi brook between Óhuta and Regéc. Here the andesite is strongly altered and cracked the fissures are filled with the mineral assemblage formed by the postvolcanic processes (*Fig. 1*).

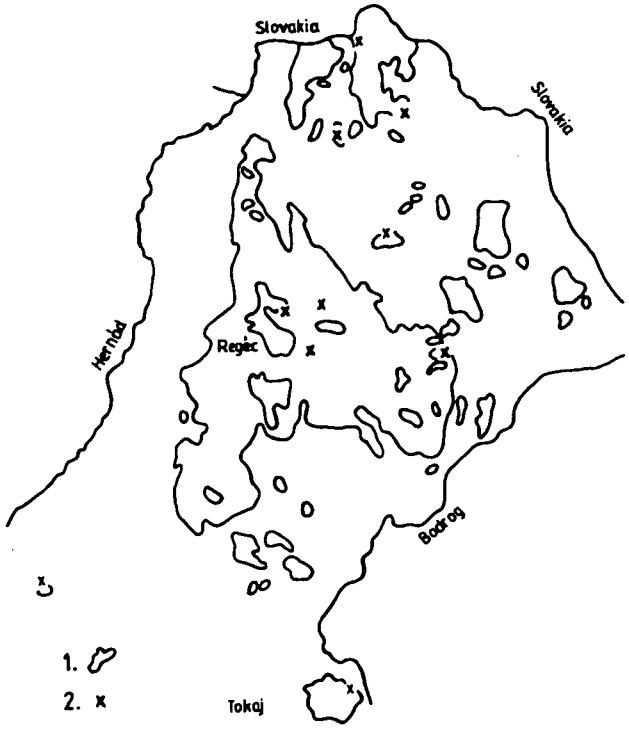


Fig. 1. Zeolite-occurrence in the intermediate volcanic rocks in the Tokaj Mts. (North-Hungary)
1. outcrops, 2. occurrences

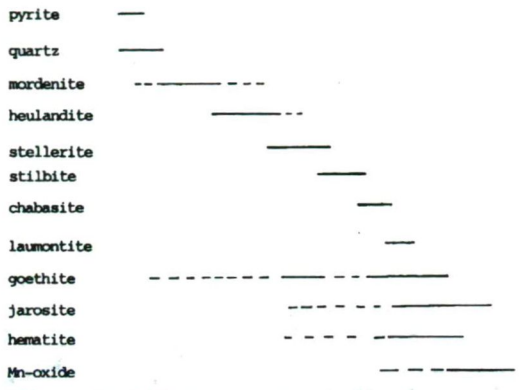
The minerals observed in order of crystallization (Table 1.): pyrite, quartz, mordenite, heulandite, stellerite, stilbite, laumontite, chabasite, hematite, goethite, jarosite.

Quartz occurs as small, short crystals or in the form of some millimetres thick chalcedony crusts. The first crystallized zeolite is mordenite which is the dominating phase of the assemblage. (To identify the zeolite phases, X-ray powder diffractometry and electron microprobe analysis were applied in each occasion.)

Modernite forms acicular crystals or found in fibrous masses. According to SEM images the individual crystals are hexagonal in cross-section and their surface is often eroded due to weathering (*Fig. 2.*).

TABLE 1.

Order of crystallization in thr zeolite-assemblage at Óhuta (Mt. Kis-Zabarla)



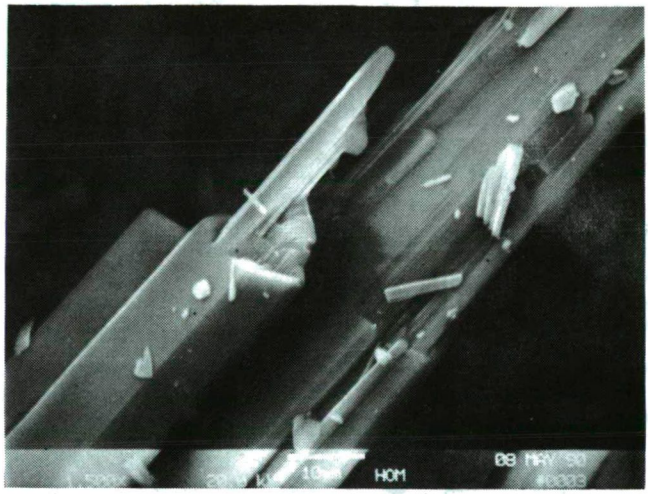


Fig. 2. Mordenite crystals, hexagonal in cross-section. Mt. Kis-Zabarla, Óhuta, SEM micrograph

Heulandite is sometimes found alone but most frequently it is associated with fibrous mordenite. It is well-developed crystals penetrated by modernite are really amazing. It is often found in clear, colourless tabular crystals reaching 1 centimetre. (100), (010), (101) forms can be observed on them.

Stellerite has coloumnar or thick tabular colourless crystals, reaching 0.5 centimetres in length, with faces of the (010), (110), (111) forms. Sometimes it is also found in flabelliform aggregates made up by 6-7 individual crystals. Stilbite is found in 2-3 centimetres large colourless or white tabular crystals with the forms (010), (110), (001).

Chabasite occurs rarely, forming colourless or white rhombohedras reaching 1 centimetre.

The latest crystallized zeolite is laumontite which has been observed twice forming milky white half centimetre large columnar crystals with the forms (110) and (201) (*Fig. 3.*).

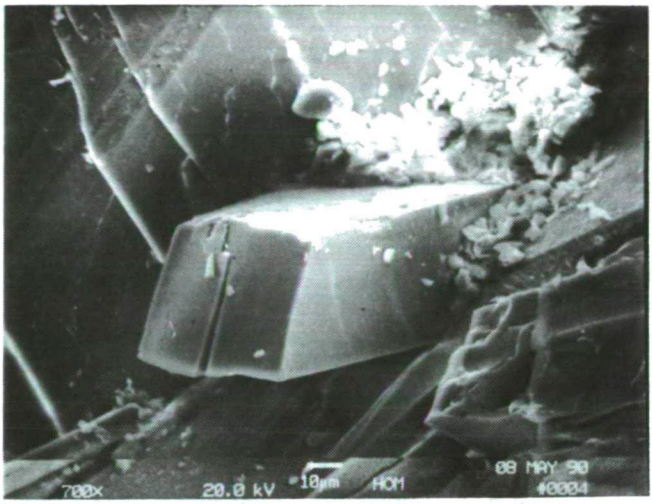


Fig. 3. SEM micrograph of a short, columnar laumontite crystal. Mt. Kis-Zabarla, Óhuta.

The zeolites are associated with disseminated pyrite, goethite, hematite, jarosite and a Manganese-oxide not identified yet.

There is also another locality near Óhuta where zeolites have been found. This is a small outcrop of a dirt track along side the Cserkő brook. Zeolites are found here in the cavities of andesite associated with celadonite and chalcedony. The most frequent species is mordenite which forms fibrous aggregates and is often found without any associating phase. There is also stellerite, occurring with mordenite (*Fig. 4.*) and sparsely chabasite is found as crystals smaller than one millimetre.

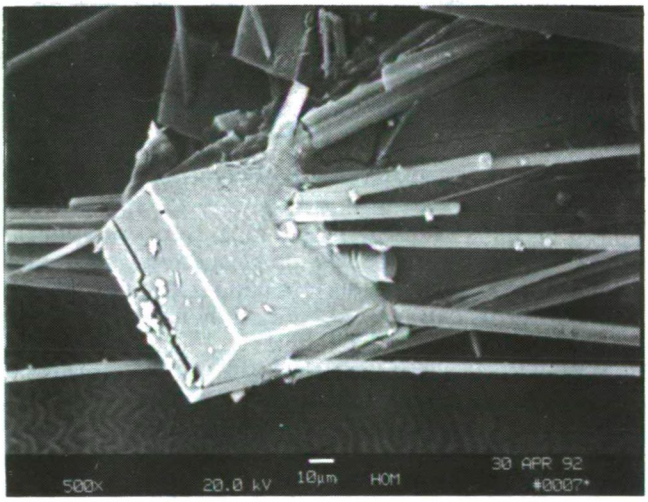


Fig. 4. Fibrous mordenite with stellerit. Cserkő-brook, Óhuta, SEM micrograph.

Regéc

Rare mineral assemblage was found in a small quarry near Mt. Torintás, north of the village. In the small cavities of andesite ferrierite and dachiardite was formed associated with pyrite hexahedras, smaller than 1 millimetre, epigene goethite, jarosite, gypsum and chalcedony.

Ferrierite is white, sometimes yellowish due to goethite and forms spherical aggregates built up by acicular crystals (*Fig. 5-6.*). This morphology is similar to that of described from Weitendorf, Austria (ZIRKL 1973). The microprobe analysis of the ferrierite from Regéc (Table 2.) suggests that it contains a lot of Mg and poor alkalies. Having regard to it is chemical composition, it is resembling to ferrierites from Monte Lake, Canada (WISE-TSCHERNICH 1976) and Philip Island, Australia (BIRCH 1989) (Table 3.). The difference between it is X-ray diffraction

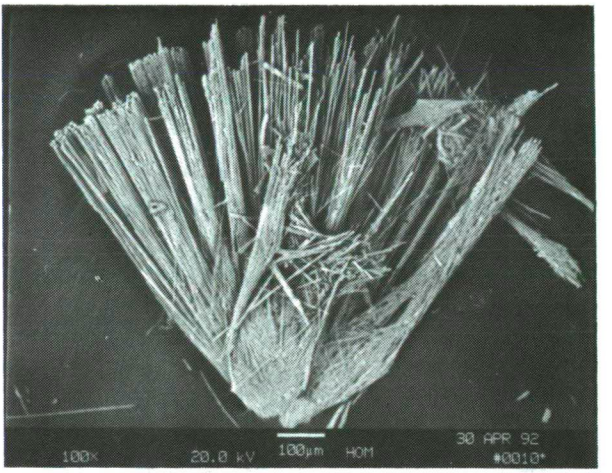


Fig. 5. Radiating aggregate of ferrierite crystals. Mt. Torintás, Regéc. SEM micrograph

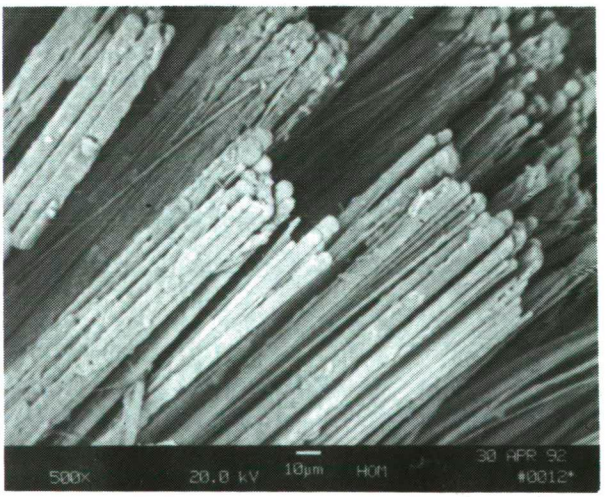


Fig. 6. Elongated ferrierite crystals. Mt. Torintás, Regéc. SEM micrograph

data on those of the right JCPDS card can possibly be caused by the difference in chemical composition (Table 4.).

TABLE 2.

Quantitave microprobe analysis of ferrierite from Regéc

mass %	1	2	3	mean
SiO_2	66.88	63.96	67.20	66.01
Al_2O_3	12.11	11.84	12.13	12.03
Fe_2O_3	0.10	0.12	0.05	0.09
MgO	3.20	3.15	3.20	3.19
CaO	1.45	1.41	1.43	1.43
Na_2O	0.12	0.12	0.12	0.12
K_2O	0.63	0.58	0.63	0.61
BaO	0.34	0.35	0.18	0.29
SrO	0.10	0.15	0.06	0.10
H_2O	n.d.	n.d.	n.d.	n.d.
Sum.	84.93	81.68	85.00	83.87
cation numbers on the basis of 72 oxygen				
Si	29.67	29.54	29.70	
Al	6.33	6.46	6.32	
Fe	0.03	0.04	0.02	
Mg	2.12	2.16	2.12	
Ca	0.69	0.70	0.68	
Na	0.10	0.11	0.10	
K	0.36	0.34	0.36	
Ba	0.06	0.04	0.03	
Sr	0.02	0.04	0.02	
Si/Al	4.7	4.6	4.7	

Anal. B. Birch (Victoria Museum, Melbourne)

TABLE 3.

X-ray powder diffraction data of ferrierite from Regéc

Ferrierit Regéc		Ferrierit JCPDS 11-429	
d(Å) obs.	int.	d(Å)	int.
9.57	49	9.61	100
7.03	12	7.00	30
6.63	9	6.61	20
5.82	16	5.84	50
4.97	7	4.96	10
4.79	2	4.80	10
4.57	3	4.58	10
4.01	55	3.99	90

TABLE 3.
(continuation)

Ferrierit Regéc		Ferrierit JCPDS 11-429	
d(Å) obs.	int.	d(Å)	int.
3.88	19	3.88	10
3.79	43	3.79	20
3.70	21	3.69	50
3.53	100	3.54	80
3.49	52	3.49	80
3.41	10	3.42	20
3.30	20	3.31	20
		3.20	10
3.15	24	3.15	30
3.07	18	3.07	30
2.97	23	2.97	30
2.90	10	2.90	20
2.72	7	2.72	20
2.58	8	2.58	30
2.48	5	2.49	30
2.42	3	2.43	20
2.36	11	2.37	40

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TABLE 4.

Chemical analyses of some ferrierites

	1	2	3	4
SiO_2	56.80	66.33	62.62	63.69
Al_2O_3	12.71	13.18	9.94	12.19
Fe_2O_3	3.29	0.27	0.58	0.04
MgO	4.12	3.41	2.61	3.16
CaO	5.52	1.39	5.78	0.53
SrO		0.34	0.18	
BaO		0.41		
Na_2O	0.27	0.50	0.13	0.58
K_2O	0.82	1.17	1.13	1.41
H_2O	14.32	13.00	13.22	
Sum.	100.79	100.00	100.16	81.80
Cat. numb. on the basis of 72 ox.				
Si	27.50	29.05	30.22	28.94
Al	7.25	6.80	5.65	6.64
Fe	1.20	0.09	0.21	0.02
Mg	2.98	2.23	1.87	2.18
Ca	0.99	0.65	0.41	0.26

TABLE 4.
(continuation)

	1	2	3	4
Sr	0.09	0.05		
Ba		0.07		
Na	0.25	0.43	0.12	0.52
K	0.51	0.65	0.70	0.83

1. Vicenza, Italy
2. Monte Lake, British Columbia, Canada
3. Weitendorf, Steiermark, (GOTTARDI-GALLI 1985)
4. Phillip Island, Victoria, Australia (BIRCH 1989)

The appearance of dachiardite is quite special (X-ray diffraction data of the dachiardite can be seen on Table 5). According to the technical literature, dachiardite is usually of acicular or radiation appearance with rare exceptions. In this locality it was observed in 2-3 millimetres large tabular crystals. SEM-images show that the crystals are commonly parallel twins, the composition surface is (001) (*Fig. 7.*) The crystals are often flattened on (001), and bushlike groups also. (001), (100), (100) forms can be observed on them.

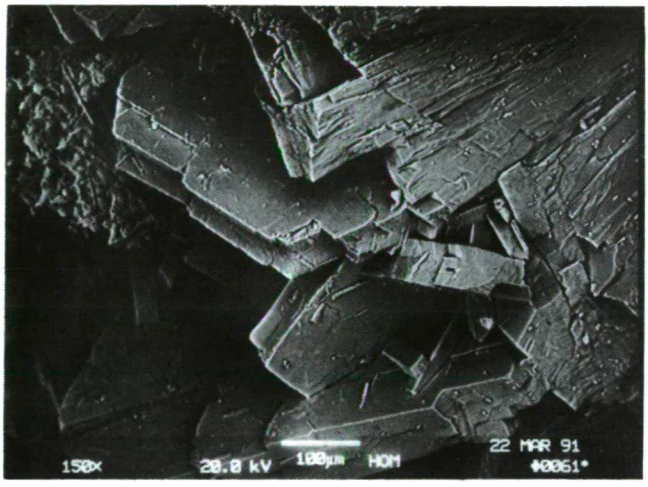

Fig. 7. Tabular twin crystals of dachiardite, Mt. Torintás, Regéc. SEM micrograph

The microprobe analyses of the samples show that they have the highest content of Ca and the lowest of Na (Table 6). High Ca and low Na content is characteristic the dachiardite variety called svetlozarite described from Bulgaria (MALEEV, 1976). Comparing the two varieties, svetlozarite contains 3.91% CaO, 0.85% NaO, dachiardite Regéc contains 5.55% CaO and 0.08% NaO. The svetlozarite is found in the Rodope Mts. in andesite-brescia in chalcedony veins. The X-ray data of dachiardite however show the best correspondence with the JCPDS card N. 30-1149 (Na-dachiardite) (Table 7.), so it needs further investigation. The associated occurrence of ferrierite and dachiardite is a rarity up to now it is described from only one assemblage formed in basalt in the USA (WISE-TSCHERNICH 1978).

TABLE 5.

X-ray powder diffraction data of the dachiardite from Regéc

Dachiardit Regéc		Dachiardit-Na JCPDS 30-1148	
d(Å)	int	d(Å)	int
9.73	6	9.77	18
8.80	100	8.86	75
6.89	6	6.92	8
5.97	4	5.96	3
4.98	34	5.00	16
4.87	8	4.88	90
		4.62	8
4.416	8	4.429	5
3.943	55	3.959	25
3.800	9	3.793	40
3.747	13	3.755	13
3.624	27	3.625	17
3.454	99	3.453	100
3.411	39	3.397	10
3.324	21	3.330	10
3.254	21	3.253	12
3.196	31	3.202	12
3.130	13	3.143	9
3.077	7	3.099	8
2.964	58	2.973	30
2.862	45	2.861	25
2.720	13	2.725	10
2.670	16	2.668	12
2.570	20	2.573	5
2.495	10	2.459	5

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TABLE 6.

Quantitative microprobe analysis of dachiardite from Regéc

mass %	1	2	mean
SiO_2	69.71	71.44	70.58
Al_2O_3	12.25	12.12	12.19
Fe_2O_3	0.07	0.10	0.09
MgO	—	—	—
CaO	5.62	5.48	5.55
Na_2O	0.12	0.05	0.08
K_2O	1.53	1.56	1.55
BaO	0.10	0.10	0.10
SrO	—	—	—

TABLE 6.
(continuation)

mass %.	1	2	mean
H_2O	n.d.	n.d.	n.d.
Sum.	89.40	90.74	90.14
Cation numbers on the basis of 48 oxygen			
Si	19.88	20.01	
Al	4.12	4.00	
Fe	0.01	0.01	
Mg	—	—	
Ca	1.72	1.65	
Na	0.07	0.03	
K	0.56	0.56	
Ba	0.01	0.01	
Sr	—	—	
Si/Al	4.8	5.0	

Anal. B. Birch (Victoria Museum, Melbourne)

TABLE 7.

Chemical analyses of some dachiardite

	1	2	3	4
SiO_2	66.15	69.58	67.38	72.41
Al_2O_3	12.91	10.26	12.65	10.15
Fe_2O_3		0.67	0.27	0.08
MgO		0.30	0.03	
CaO	3.88	3.91	0.51	1.78
Na_2O	2.57	0.85	5.15	3.04
K_2O	0.95	3.00	0.97	1.30
H_2O	14.42	10.94	12.91	
Sum.	99.88	99.51	100.00	
Cation numbers on the basis of 80 oxygen				
Si	19.50	20.19	19.76	20.62
Al	4.55	3.51	4.37	3.40
Fe		0.15	0.06	0.02
Mg		0.13	0.01	
Ca	1.24	1.22	0.16	0.55
Sr	0.02			
Ba			0.02	
Na	1.49	0.48	2.93	1.67
K	0.36	1.11	0.36	0.47
Cs	0.12			
H_2O	14.39	10.59	12.63	

1, Elba, Italy
2, "svetlozarit", Rodope, Bulgaria
3, Tsugawa, Japán
4, Altoona, Washington, USA (GOTTARDI-GALLI 1985).

Other occurrences

Stellerite and heulandite crystals associated with chalcite and pyrite in small cavities of rock were found in a quarry outcrop pyroxene-dacite in rhyolitic area near Rostalló, Kishuta. Stellerite crystals making up bunches are also common (*Fig. 8.*).

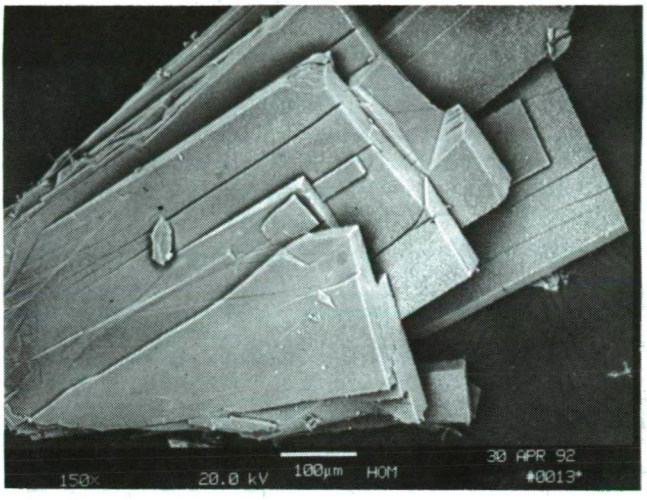

Fig. 8. Tabular stellerite forming tufted bunches. Rostalló, Kishuta. SEM micrograph

Stellerite, dachiardite and heulandite crystals can be found with pyrite, quartz, goethite, jarosite and gypsum in small cavities in an outcropt Mt. Kis-Sertés near Regéc.

Heulandite was found on Mt. Magos Hollóháza, Patkóbánya, near Tokaj and at Picpangos-spring, near Makkoshotyka. All these outcrops produced heulandite in aggregates of thick, tabular crystals without accesory minerals.

CONCLUSIONS

The zeolite assemblages in the intermadiate rocks of the Tokaj Mts. were formed due to in the course of postvolcanic processes by comparison with analogies from the technical literature (GOTTARDI-GALLI 1985). Two types of assemblages can be observed. One of them is caracterized by quartz (and chalcedony) as a first crystallized phase. They were followed by mordenite and later formed other zeolites in smaller quantities. This kind of assemblages are found particularly in the central areas to decompose due to weathering processes. JÁNOSSY *et al.* (1987) described a similar paragenesis and formation.

The other type is a poor in species. The formation of minerals follows the scheme that begins with the crystallization of quartz and chalcedony later precipitating one or two species of zeolites, most commonly heulandite, sometimes stellerite. The ferrierite-dachiardite bearing paragenesis at Regéc belongs to this better type, though due to the difference in the composition of the hydrothermal solutions not the ordinary zeolites were formed.

To summarize the results, we can accept that the presence of zeolite assemblages in the intermediate rocks of the area in question is familiar and is part of the postvolcanic processes.

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PIEDMONT, RECHNITZ AND MELIATA ZONE: A PETROGRAPHIC-GEOCHEMICAL COMPARISON OF METAMORPHIC OPHIOLITES OF THE ALP-CARPATHIAN SYSTEM

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ABSTRACT

In spite of the polyphase metamorphic history and intense deformation there are a lot of ultramafic, gabbro, and plagiogranite, mafic dykes, volcanic and deep-sea sediment occurrences of Mesozoic age in the Alp-Carpathian system. Our petrographic-geochemical comparison of these rock-types focuses on two zones of the Alpine belt: the Western Alps (Piedmont) and the Eastern Penninic Alps (Rechnitz) and one zone of the Carpathians: the North Hungarian and South Slovakian Meliaticum. The first zone is characterized by the abundant relics of high pressure metamorphism of Cretaceous age which survived the greenschist facies metamorphism of Tertiary age, while the two latter show relatively limited relics of the (eoalpine) glaucophane schist facies conditions.

The special interest of a comparison derives from the fact that the first two zones form a a typical Jurrasic-Cretaceous ophiolite nappe occurring in the west- and easternmost zones of the Alpine belt. These two zones display a genetic similarity. On the other hand, the third zone is considered as a Triassic-Jurassic incomplete ophiolite suite occurring in the present Innermost Carpathian belt.

Based on detailed field and laboratory studies we compare the three zones in terms of their upper mantle rocks (serpentinite and pyroxene-rich chlorite schist), mafic plutonic suite (coarse grained eclogite, glaucophane schist- and greenschist-facies metagabbro and plagio-granite), volcanic suite (fine grained eclogite, glaucophane schist- and greenschist-facies metavolcanic) and sedimentary cover (radiolarite, ophiolitic breccia, ophicarbonate, calcschist, micaschist and quartzite).

The lithologies of the two zones of the Alps are relatively similar and comparable to the Western Alps metaophiolites (N- and T-type MORB volcanics, range of plutonic rocks from primitive Mg-Al rich gabbros to late fractionated plagiogranites, predominance of depleted lherzolites and/or harzburgites among the ultramafics). On the other hand, the lithology of the third zone (Meliaticum) of the Innermost Carpathians is relatively different from the ophiolite zones of Alpine belt (E- and T-type MORB, spilitic volcanics, albite-rich gabbros, dominant harzburgitic and minor lherzolitic ultramafics) which may exclude a close genetic connection between the two belts. Moreover, the volcanic (and probably the plutonic) rocks of the Alps show a much greater compositional variability, suggesting a relatively slow spreading rate. In contrast, the Meliata volcanics display almost a limited compositional variation and sometimes spilitic pillow lavas, suggesting a relatively fast spreading rate with occasional interaction with seawater.

There are minor but interesting similarities among these ophiolites. All of them are high-Ti type, have tholeiitic affinitites with a tendency to "transitional" MORB character. All these ophiolite-sequences show a progressive enrichment of both rare earth and incompatible elements starting from ultramafics toward the effusive members as a result of magmatic differentiation. However, we found a Ba enrichment in all volcanic members, which can be derived from the deep-sea sediments covering the volcanics. All of these ophiolites have undergone some oceanic alterations and the complex Alpine regional metamorphic events, and mixed or covered by marine sediments.

In conclusion, the comparison between the Alpine and Carpathian ophiolites lead to assign a peculiar significance among them (as previous) in relation to the various tectonic events dominating in the Jurassic-Cretaceous Piedmont-Ligurian basin and in the Triassic-Jurassic Vardar basin.

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INTRODUCTION

Both the Piedmont and Rechnitz zones belong to the Penninic domain ophiolite in the Western and Eastern Alps respectively, whereas the Meliata zone appears to be related to the Pannonian basin in the Innermost Western Carpathians and/or Inner Dinarids. In the Piedmont zone, much more tectonometamorphic complications and more typical metaophiolite sequences have been developed with respect to that found in the Rechnitz and Meliata zones. These zones together form a west-east geotraverse of about 1100 km from Montgenévre to South Slovakia (*Fig. 1*). Although an increasing number of studies with emphasis on petrogenesis and geochemical aspects are available (Western Alps: LOMBARDO and POGNANTE 1982; LOMBARDO *et al.* 1978; LEWIS and SMEWING 1980; DAL PIAZ *et al.* 1980, 1981; BERTRAND *et al.* 1987; PFEIFER *et al.* 1989; Rechnitz zone: PAHR and KOLLER 1980; KOLLER 1985; HÖCK and KOLLER 1989; Meliata zone: HOVORKA 1983, 1985; RÉTI 1986; ÁRKAI and KOVÁCS 1986; KUBOVICS *et al.* 1990), the comparison of the three regions are often neglected or not explicitely considered in the framework of the evolution of the Alp-Carpathian system in the Western Tethys.

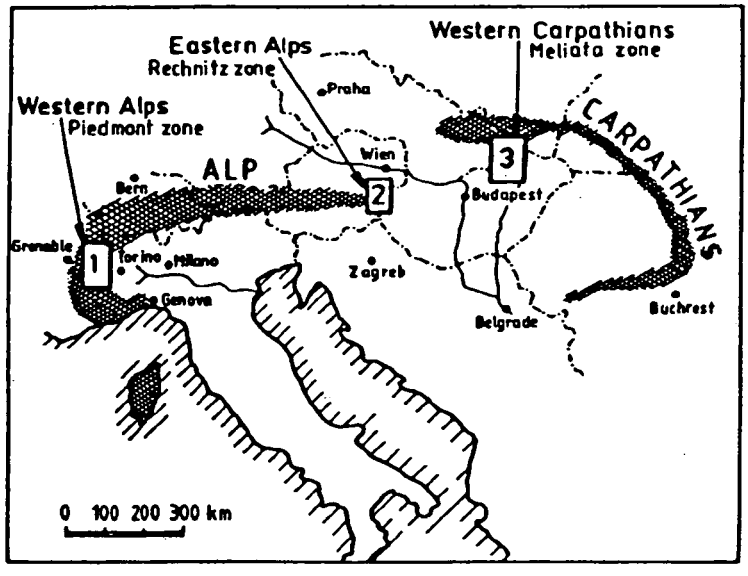

Fig. 1. Location of the three regions studied.

ROCK TYPES

Most of the common members of the Western Tethys ophiolite suites have been identified along the whole traverse explored here, independently of the metamorphic grade. Table 1 summarizes the different rock types encountered. The ultramafic, gabbroic and volcanic rocks are sporadically distributed and occurred mostly in separated tectonic slices.

Ultramafic rocks, in the Western Alps region typically occur as large (some km^2) sheets, sometimes containing a few metre thick lenses and boudins of gabbros

TABLE 1.

Metamorphic analogues of the different series of an ophiolite suite: rock types as a function of metamorphic grade (typical HP-facies to greenschist facies in the Western Alps, greenschist facies with blueschist relics in the Eastern Alps and slightly greenschist with blueschist relics in the western Carpathians)

ORIGINAL OPHIOLITE MEMBER	METAMORPHIC EQUIVALENT		
	western region Western Alps (Piedmont zone)	middle region Eastern Alps (Rechnitz zone)	eastern region Western Carpathians (Meliata zone)
sediments			
- calcareous sed.	- phengite-, chloritoid-bearing calcschists; cpx-bearing marbles; garnet-, tourmaline-bearing micaschists	- phyllite; marbles; albite-bearing chlorite schists	- crystalline limestones; phyllites
- siliceous sed.	- quartzites	- quartzites	- argillites; radiolarites
- ophiolitic detritus	- serpentinite breccias; ophicalcites	- ophicalcites	- serpentinite breccias; ophicalcites
volcanic rocks			
- acidic differentiates	- no equivalent found	- no equivalent found	- keratophyres; qtz-porphyres
- hyaloclastites (partially pillows)	- glaucophanites; prasinites; ovardites; (fine-grained)	- prasinites; tuffites; (fine-grained); glaucophane metabas.	- glaucophane metabasalts
- basalts s.l.	- omphacite-rich eclogites; cpx-plag-metabasalts	- greenschists (fine-grained)	- cpx-plag-metabasalts; spilites (fine-grained lava flows and pillows)
dykes	- eclogites (fine-grained); basalts and dolerites (porphyritic) usually cutting gabbros	- greenschists (fine-grained)	- microgabbros; dolerites (variable grain-size)
	- plagiogranites (albitites)	- plagiogranites (blueschists)	- no equivalent found
mafic plutonites			
- Mg-Al gabbros (normal gabbros)	- greenschist facies metagabbros; tourmaline cummingtonite metagabbros; smaragdite meta-gabbros; chloritized metagabbros; eclogites (coarse-grained)	- sheared cpx-metagabbros; flaser gabbros	- normal gabbros; spilitic metagabbros (medium grained)
- intermediate gabbros	- cpx-metagabbros		
- Fe-Ti gabbros	- eclogitic metagabbros; glaucophane-metagabbros; ferrogabbros	- glaucophane metagabbros; ferrogabbros	- ferrogabbros (?)
ultramafic rocks			
- dunites/harzburgites		- antigorite serpentinites	- ol-opx-cpx-sp-serpentinites
- lherzolites	- diopside-Ti-clinohumite-bearing serpentinites; trem-cpx serpentinites	- lizardite-chrysotile serp.	- lizardite-chrysotile serpentinites
- pyroxenites	- cpx-tremolite-, cpx-garnet-chlorite rocks	- cpx-pump-garnet chlorite rocks	
- metasomatic ultramafic rocks	- chlorite schists, talc serpentinites	- talc-chlorite schists; trem-chlorite-schists; talc serp.	
Ca-rich mafic rocks	- garnet-diopside-epidote-Fe-Ti spinel rocks often cutting serpentinites	- cpx-garnet-chlorite-pump-plag rocks	- garnet-cpx-chlorite-serpentine rocks (?)

and eclogites. Those ultramafics also occurring as some tens of metres within the mafic rocks. In the Inner Western Carpathians (Meliata zone) the ultramafics are the most abundant rocks, occurring as bodies of various size (up to a few km^2) in an evaporitic melange (RÉTI 1985). In all the three regions, the exact origin of the rocks (tectonic mantle peridotites, or cumulate series at the crustal mafic plutonic suite) cannot be determined due to the intensive deformation (dismembering and folding). All the ultramafics are metamorphic, intensively serpentinized and reflect the Late Mesozoic metamorphism. Only a few relics spinel, clinopyroxene and very scarce olivine were preserved in these rocks. In the marginal parts of ultramafic masses, thin layers or dykes (few dm thick) of tremolite-chlorite schists and/or metapyroxenites can be observed.

Mafic plutonic rocks, in the Western and Eastern Alps scarce relics of magmatic stage have survived the polyphase metamorphism. The gabbros are systematically coarser grained (flaser structure at cm-scale) and are mostly lighter green colour than the volcanic rocks. The best preserved gabbros in the Western Alps are found in Montgenévre containing cpx ranges from augite to salite (BERTRAND *et al.* 1987). In both Western and Eastern Alps zone, the gabbroic sequence include primitive members (Mg-Al gabbros) and more differentiated Fe-Ti gabbros and plagiogranites. On the other hand, in Meliata zone, the gabbroic rocks are very rare and different from the former gabbros. Those are moderately spilitized and overprinted by greenschist facies condition.

Ca-rich mafic rocks (metarodingites), in all the three regions garnet-cpx inclusions of metre-scale with rim of chlorite-amphibole rocks occur in close relationship to ultramafic masses. These metarodingites are strongly boudinaged, but in some places their original gabbroic content can still be recognized in the Piedmont and Rechnitz zone (DAL PIAZ *et al.* 1980; KOLLER 1985). In Meliata zone rodingitic contacts of ultramafics with accompanying sedimentary structure also occur (HOVORKA 1985).

Dykes, outcrops of subvolcanic rocks are very scarce and are often found as a few cm to several dm thick dykes crosscutting the gabbros and more rarely ultramafics. Generally, a proper sheeted dykes complex originally situated between the plutonic and volcanic suite can not be found in either region (DIETRICH 1980; LOMBARDO and POGNANTE 1982; KOLLER 1985; KUBOVICS *et al.* 1990).

Volcanic rocks, on the Piedmont and Rechnitz zone. Original pillow structure are mostly preserved (OBERHANSLI 1980; LEWIS and SMEWING 1980; BERTRAND *et al.* 1987). In some of these volcanic rocks, original structure as well as the primary mineral were disappeared. However, in the Meliata volcanics, the pillow structure is frequently preserved (HOVORKA and SPISIAK 1988).

Sedimentary rocks, in the Piedmont zone thick piles of calcschists together with marble, micaschist, quartzite +/- ophicarbonate and ophiolitic breccias occur. A large part of the calc-schists appear to be Cretaceous in age (MARTHALER 1984; LAGABRIELLE 1987). Most of the sediments of the Piedmont zone contain detritus of both oceanic and continental crust (SARTORI 1987; STECK 1989). These sediments are usually mixed with resedimented volcanic materials (layers of eclogite and greenschist). In the Rechnitz, the sediments are dominantly consisting of phyllitic rocks and are equivalent to oceanic sediments (PAHR and KOLLER 1980). In Meliata zone, the sediments are mainly represented by Middle Triassic radiolarite (KOZUR and RÉTI 1986), argillite and siltstone and are considered as oceanic sediments (KOVÁCS 1984; RÉTI 1987).

Oceanic metamorphism, in the Piedmont zone, produced only local development of brown- and green hornblende. In the Rechnitz, oceanic metamorphism is characterized by barroisite and Mg-hornblende with temperature of formation lower than 750 °C and pressure about 1 kbar (KOLLER 1985).

Cretaceous (Early-Alpine) high pressure facies, in the Piedmont zone, the Early-Alpine event is polystadial with an early blueschist facies followed by an eclogitic facies for which maximum conditions of 470—450 °C and 10 kbar for Monviso ophiolite (MONVISO 1980) and 600—500 °C and about 24—18 kbar for regions around Zermatt and Saas Fee have been established (OBERHANSLI 1980, 1986; MEYER 1983; BARNICOAT and FRY 1986; GANGUIN 1986, 1988). These eclogites were pervasive and produced mineral assemblages consisting of Na-pyroxene, garnet, epidote, rutile +/- Mg-chlorite and talc. Blueschist facies have been recorded in all three zones but with different strength, condition and abundance. They include Na-pyroxene, Na-amphibole, albite +/- garnet and white mica. This paragenetic assemblage appears to yield temperatures of 450, 370—330, 450—380 °C and pressures of 8—7, 8—6 and 7 kbar for blueschists from Monviso in the Piedmont (MONVISO 1980), Rechnitz (KOLLER 1985) and Meliata zone (ÁRKAI and KOVÁCS, 1986; HOVORKA, 1990, pers. comm.) respectively.

Tertiary (Meso-Alpine) greenschist facies, the first retrograde overprint in the Piedmont zone areas produced typical barroisitic and pargasitic amphibole and albite (GANGUIN 1988). This stage correspond to be barroisitic blueschist facies of MONVISO (1980). Their conditions of formation is estimated at about 500—450 °C and 10—6 kbar (PFEIFER *et al.* 1989). The second retrograde overprint of early-Alpine parageneses in the Alps (Western and Eastern Alps) produced mineral assemblage including albite, Fe-chlorite, Fe-epidote, actinolite and titanite which indicate greenschist facies conditions. In the Piedmont, these occur at about 400 or 500—450 °C and 5—3 kbar (MONVISO 1980; PFEIFER *et al.* 1989). The greenschist facies parageneses of both Rechnitz and Meliata zones seem to be characterized by the same conditions of temperature and pressure showing 430—390 °C and about 3 kbar (KOLLER 1985; ÁRKAI and KOVÁCS 1986). Furthermore, in addition to greenschist; prehnite-pumpellyite facies is also recorded in the Meliata zone (Table 2.).

Metamorphism of ultramafic rocks, because ultramafic rocks are little sensitive to high pressure, the early- and meso-Alpine mineral associations are difficult to distinguish. They were frequently deformed and serpentinized. However, fragments already serpentinized (in the oceanic environment) more easily crystallized to metamorphic serpentinite (PFEIFER *et al.* 1989).

BULK ROCK COMPOSITION

General chemical feature, in a systematic survey of the mentioned W—E geotraverse around 46 mafic and 24 ultramafic samples have been analyzed from the three ophiolite regions. The bulk rock (major, trace and rare earth element) analyses of all rock type are given in ABDEL-KARIM (1990) and KUBOVICS *et al.* (in this volume). The ultramafic rocks of all three regions have similar values of Mg, Mn and Sc. Furthermore, in both Piedmont and Meliata zone, these ultramafic show similar values of Al, Fe, K, Y and Zr, too. The gabbroic rocks of the Piedmont and Rechnitz zone are correlated each other in their major and trace element compositions, showing similar values of Si, Ti and Mn. However, higher values

TABLE 2.

Possible metamorphic facies conditions and their mineral assemblages distributions of ophiolite suites from the Alp-Carpathian system

METAMORPHIC FACIES	WESTERN ALPS	EASTERN ALPS	WESTERN CARPATHIANS
	Piedmont zone	Rechnitz zone	Meliata zone
1. Oceanic event			
Pressure (kb)[+]		≤1	
Temperature (°C)[+]		< 750 (?)	
Min. assemblages:			
barroisite	———— - -	———— - -	
Mg-hornblende	———— - -	———— - -	
2. Eclogite facies			
Pressure (kb)[+]	10; 24-18		
Temperature (°C)[+]	470-450; 600-550		
Min. assemblages:			
Na-pyroxene	————————	— -(?)	
garnet	————————	— -(?)	
epidote	————————		
rutile	————————		
± Mg-chlorite	————————		
± talc	————————	———— -	
3. Blueschist facies			
Pressure (kb)[+]			
Temperature (°C)[+]	8-7	8-6	7
Min. assemblages:	450	370-330	330-450
Na-pyroxene	————————	————————	
Na-amphibole	————————	————————	————————
albite	————————	————————	————————
± white mica (phengit)	————————	———— -	— -(?)
± garnet	————————	——	———— -
4. Barroisitic blueschist facies			
Pressure (kb)[+]	10-6		
Temperature (°C)[+]	500-450		
Min. assemblages:			
albite	————————		
barroisite (blue-green hornblende)	————————	— -	
5. Greenschist facies			
Pressure (kb)[+]			
Temperature (°C)[+]	5-3	3	3
Min. assemblages:	400; 500-450	430-390	350
albite	————————	————————	————————
Fe-chlorite	————————	————————	————————
Fe-epidote	————————	————————	————————
amphibole (trem.-actin.)	————————	————————	———————— -
titanite	————————	————————	———— -
± biotite (stilpnom. ?)	———————— -	———— - -	— -

[+]Data were compiled after MONVISO 1980; KOLLER 1985; PFEIFER *et al.* 1989; ÁRKAI and KOVÁCS 1985; HOVORKA 1990 (pers. comm.)

of Mg, Sr, Y, Zr, Sc but lower Ba and Rb contents of the Piedmont zone when compared with Rechnitz are observed. On the other hand, the gabbroid rocks of Meliata zone are mostly different from that of the two former zones. The glaucophane schist facies metavolcanics of the Piedmont and Meliata zone are similar in most major element compositions however, significant higher values of Rb, Cr, Hf and light REE are associated with Meliata indicating their strong tendencies with the E-type MORB. The greenschist facies metavolcanics of both Rechnitz and Meliata metavolcanics are enriched in Ba, Sr, Nb, REE but lower Rb, Y and Cr when compared with the two other zone probably due to their E-type MORB characters.

Mafic (volcanic and plutonic) and ultramafic rocks are in general well discriminated on a number of major and trace element diagrams. In AFM-diagram (*Fig. 2*), the ultramafic rocks of the three regions fall within the metamorphic peridotite (COLEMAN 1977) and mafic-ultramafic cumulate field drawn by STRONG and MALPAS (1975). However, the chlorite schists (including chloritites) of the Rechnitz zone locate on the F-M side showing an enrichment of FeO^t in comparison with the ultramafics. Comparing the Meliata zone with the other two regions, true Alpine gabbros (Mg-Al rich and Fe-Ti rich gabbros) are absent in the former, that may be attributed to the enrichment of these gabbros in Ca instead of Mg and the effect of spilitic process. In the Piedmont zone, the altered intermediate gabbros and most volcanic rocks are enriched in alkalis may be due to variable element mobilization during the different Alpine metamorphic events. All the Alpine gabbros fall into the oceanic gabbro field (BONATTI *et al.* 1971; THOMPSON 1973; PRINZ *et al.* 1976; CAYTROUGH 1979). The albitites (in Piedmont) and keratophyres (in Meliata) show a rather calc-alkaline tendency. However, the ophiolites of the three regions show tholeiitic character with a marked Fe-Ti enrichment trend.

REE patterns of the mafic-ultramafic rocks of all three regions normalized to the chondrite values given by NAKAMURA (1974) are shown in *Fig. 3.* The ultramafic rocks of both the Piedmont and Meliata zone exhibit the same compositional trend, showing slight light REE-depletion. However, LREE-depletion and HREE enrichment are more distinct in the former zone than in the latter. The serpentinite, rodingite and chlorite schist of the Rechnitz zone have relatively the same trend. Comparing the gabbroic rocks of Piedmont zone with that from Rechnitz wide compositional range and a more abundant REE-content are recorded with the former. The REE patterns of the Meliata gabbros have the narrowest compositional range and is significantly different from that the other two zones. The albitites from Montgenévre (in the Piedmont zone) and the plagiogranites (blueschists) from Rechnitz zone are similar in REE patterns. The mafic volcanic rocks of the Piedmont and Rechnitz zone are characterized by analogous REE-abundances showing a N-MORB character that different from Meliata zone which exhibit an E-MORB affinity. In the all three regions, a progressive REE-enrichment from ultramafics through gabbroics and volcanics to plagiogranites occur. These data are consistent with the results given by the patterns of incompatible trace and minor elements in normalized spiderdiagrams (after THOMPSON *et al.* 1984; *Fig. 4*) of the same rocks from the three regions. The most of the REE and incompatible elements show typical magmatic rock trends which confirm that the original bulk rock-trace element compositions can be preserved even during the different stage of the metamorphism.

Ultramafic rocks, all the three regions, the trends observed during petrographic inspection are confirmed by the bulk rock chemistry: Al- and Ca-poor dunitic and harzburgitic compositions are limited to strongly deformed, and probably Ca-depleted serpentinites; some rocks contain typical Al_2O_3 contents of lherzolite,

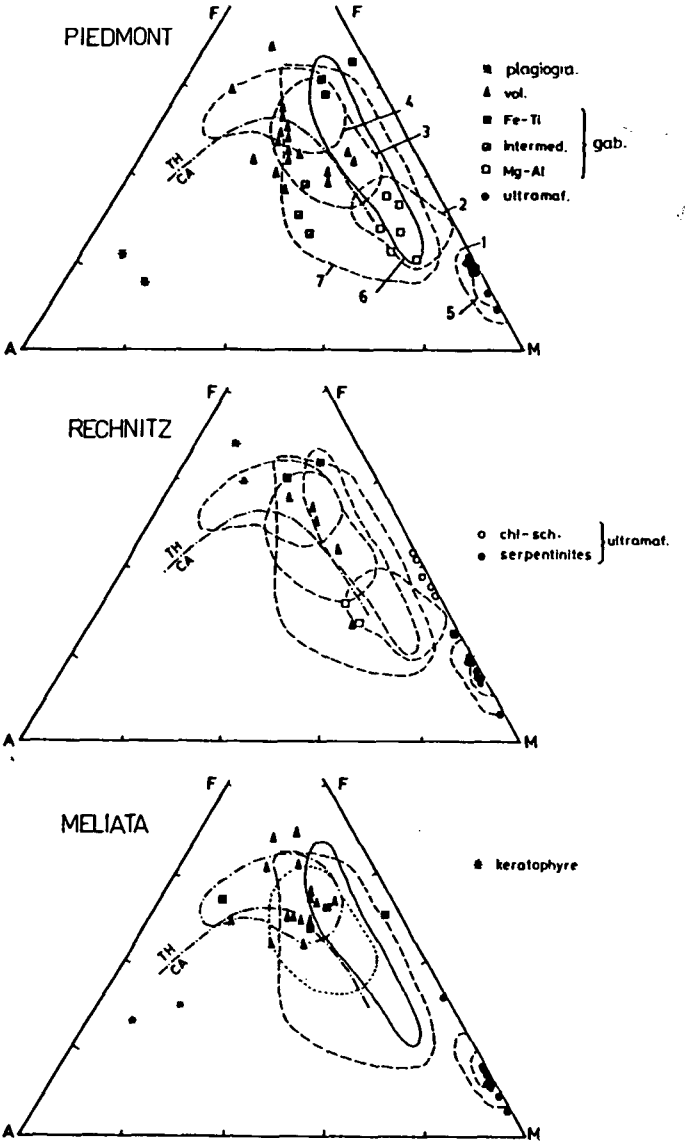

Fig. 2. Mafic-ultramafic rock compositions in the AFM diagram (in weight %) as a function of the different regions (A: K_2O+Na_2O, M: MgO, F: FeO^t). Fields-1: mafic-ultramafic cumulates, 2: gabbros, 3: sheeted dykes, 4: lavas (after STRONG and MALPAS 1975), 5: metamorphic peridotites (COLEMAN 1977), 6: gabbros from Mid Atlantic Ridge (BONATTI *et al.* 1971; THOMPSON 1973; PRINZ *et al.* 1976). 7: gabbros from Mid Cayman Ridge (CAYTROUGH 1979).

harzburgite and rare pyroxenites, comparable to other western Tethyan ultramafics (BECCALUVA *et al.* 1984; POGNANTE *et al.* 1986; KUBOVICS and ABDEL-KARIM 1990).

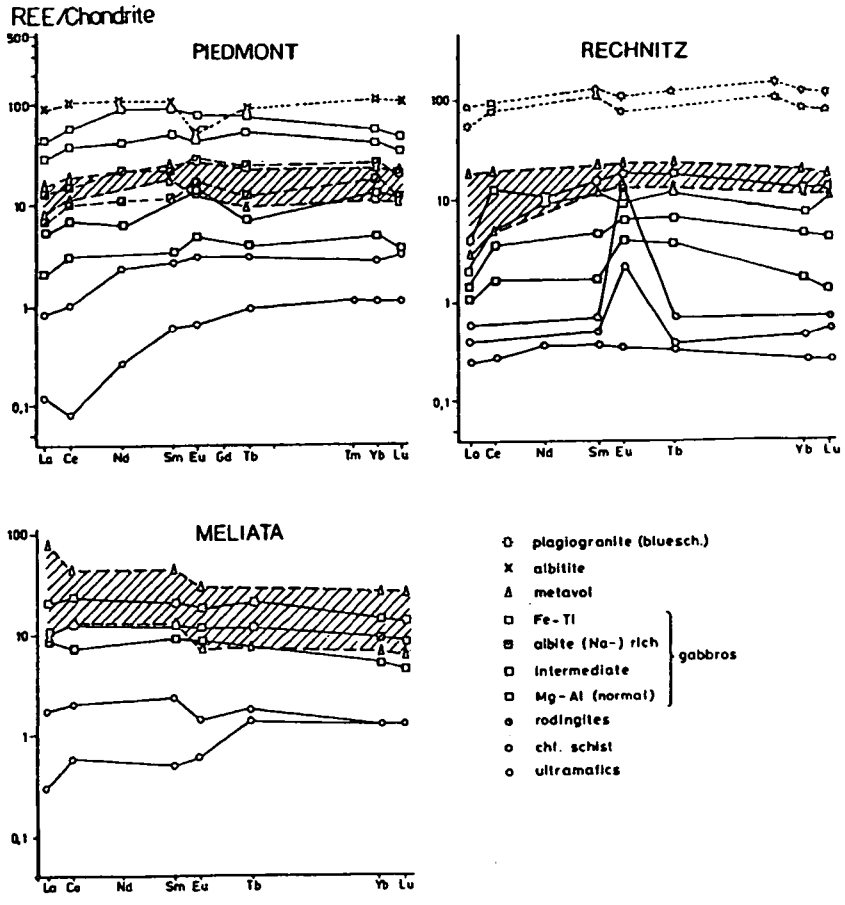

Fig. 3. Rare earth element patterns of the ophiolite suites from the three regions, normalized to chondrite values of NAKAMURA (1974) and LEWIS and SMEWING (1980). Sources of data from Piedmont zone (LEWIS and SMEWING 1980; PICCARDO *et al.* 1988 c,d; our unpublished data) and from Rechnitz zone (KOLLER 1985; our unpublished data).

Pattern of 3-d transition elements normalized to "primitive" mantle estimated of JAGOUTZ *et al.* (1979) of the ultramafic rocks of all the three regions desplay compositions comparable to each other (*Fig. 5*). In detail, there are some remarkable differences between them. Comparing the Rechnitz zone with the Piedmont and Meliata zone, it is significant higher Ti and Fe and lower Ni values of the former. However some ultramafics from Piedmont zone show lower Ti and higher Co values with respect to other two. Many ultramafic rocks of Meliata zone are depleted in V and Mn and enriched in Co and Ni contents, as compared with the Rechnitz zone.

Plutonic rocks, because magmatic mineral phases have been preserved rarely, protoliths of metagabbroic rocks are identified either based on their coarse grained texture or on major and trace element patterns (*Fig. 2*). Three groups can be recognized: (1) Mg-Al rich metagabbros (normal gabbros) with higher MgO,

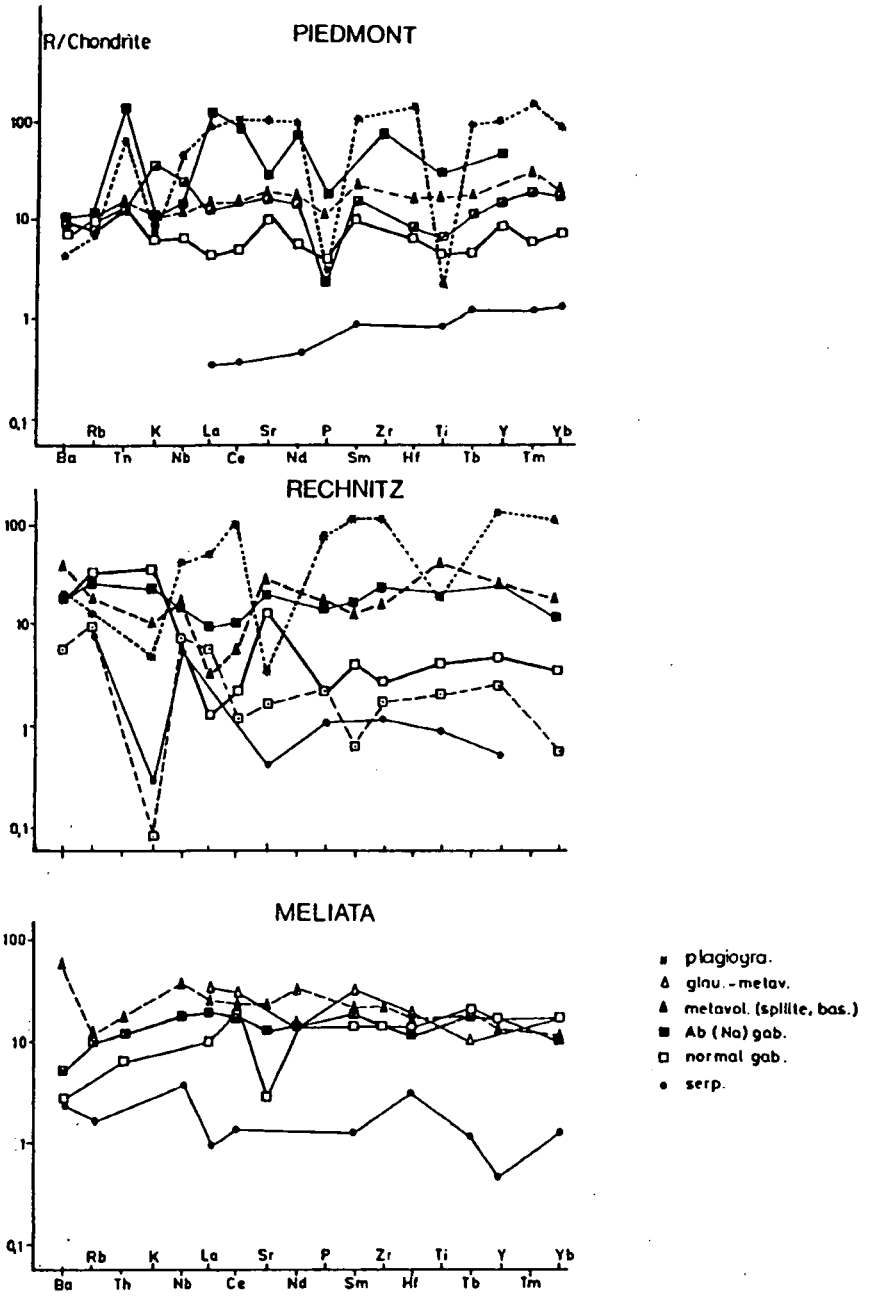

Fig. 4. Composition of the common trace element contents (as ppm) of ophiolite suites from the three regions normalized to chondrite values of THOMPSON *et al.* (1984). Sources of data from Piedmont ultramafics (OTTONELLO *et al.* 1984) and from Rechnitz (KOLLER 1985; our unpublished data).

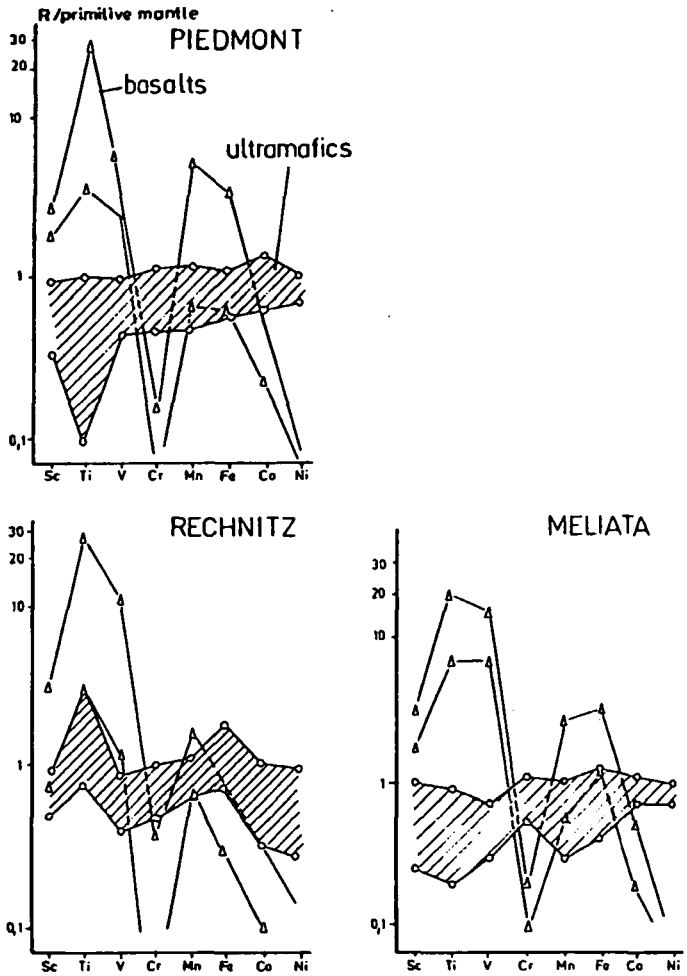

Fig. 5. 3-d transition element composition of ultramafic and volcanic rocks of the three regions, normalized to the primitive mantle composition (JAGOUTZ *et al.* 1979). Sources of data from Piedmont zone (POGNANTE *et al.* 1986), Rechnitz zone (KOLLER 1985) and from Meliata zone (HOVORKA 1985; our unpublished data).

Al_2O_3 and Cr and with low FeO^t/FeO^t+MgO values; (2) intermediate metagabbros with an intermediate value of FeO^t, FeO^t/FeO^t+MgO, (about 0.3—0.5) and Y, corresponding to plagioclase-pyroxene-gabbros and (3) Fe-Ti group of ferrogabbros and rarely glaucophane gabbros, with very high FeO^t, TiO_2, FeO^t/FeO^t+MgO (> 0.6) and with low SiO_2. The three groups of gabbros considered as typical stages of tholeiitic differentiation. The gabbroic rocks of the Alps (from Piedmont and Rechnitz zone) characterized by a progressive enrichment of REE and incompatible elements (*Fig. 3—4*) during their differentiation from Mg-Al gabbro to Fe-Ti gabbro and plagiogranite. On the other hand, very little progressive enrichment trend from Ca-rich (normal) gabbros to Na-rich (albite) gabbros have been found.

The *volcanic rocks* of all three zones display composition comparable to present-day mid-ocean ridge or marginal basin basalt of constructive oceanic plate margins (MORB, *Fig. 6—7*). However there are a few minor but interesting differences among the three regions. In the REE pattern (*Fig. 3*) and Hf-Th-Ta diagram (WOOD *et al.* 1979; *Fig. 6*) the Piedmont basalts show closer similarities with normal and transitional (N- and T-) MORB; the Rechnitz basalts are typically N-MORB (KOLLER 1985), while the Meliata basalts exhibit closer affinities with enriched and transitional (E- and T-) MORB.

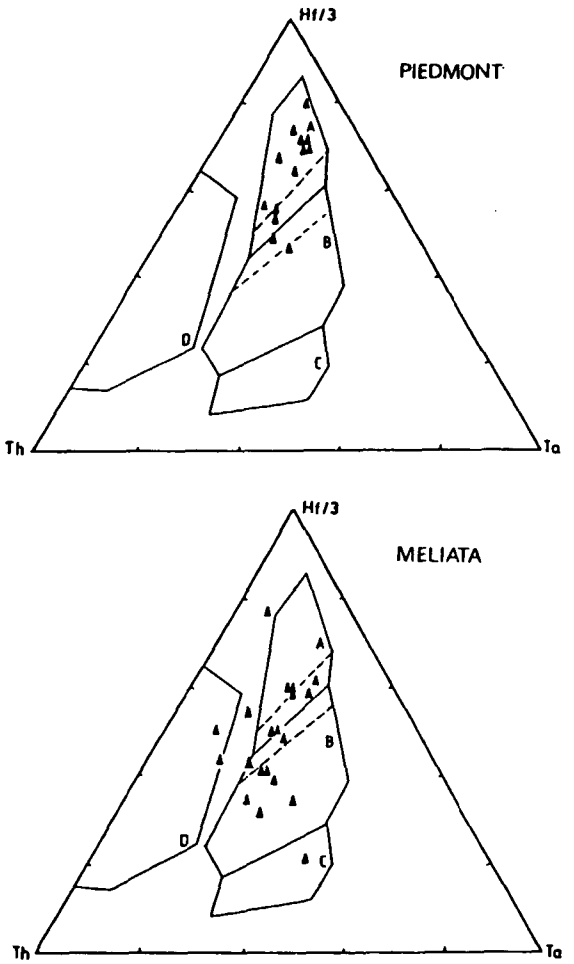

Fig. 6. Th-Hf/3-Ta diagram of basaltic rocks for the Piedmont and Meliata zones. A: N-type MORB; B: E-type MORB; C: alkalic within-plate basalts; D: magmas at destructive plate margins (after WOOD *et al.* 1979a). Sources of data: Piedmont basalts from LEWIS and SMEWING (1980) and our unpublished data.

The volcanic rocks of both Piedmont and Rechnitz zone plot quite well on a straight line passing through the origin of diagrams relating two hydromagmatophile elements (WOOD *et al.* 1979a) indicating that they are related by crystal

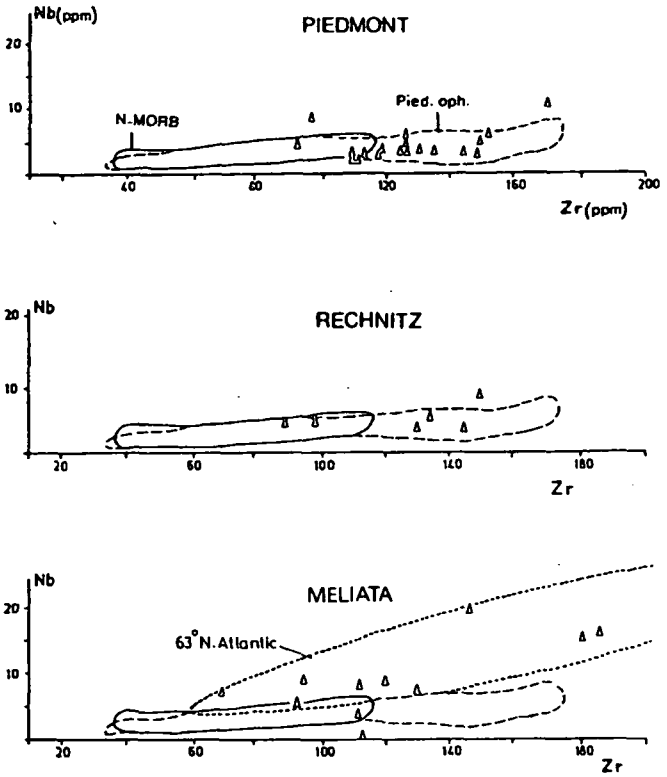

Fig. 7. Nb-Zr diagram (in ppm) of basaltic rocks for the three regions. The field of Piedmont basalts (LOMBARDO *et al.* 1978; LEWIS and SMEWING 1980; DAL PIAZ *et al.* 1981; BECCALUVA *et al.* 1984b; POGNANTE *et al.* 1982,1985). The fields of some MORB and some basalts sampled at different latitudes in the North Atlantic (PEARCE and CANN 1973; SUN *et al.* 1979; WOOD *et al.* 1979; LE ROEX *et al.* 1982) are shown for comparison. Sources of data from Piedmont zone (LEWIS and SMEWING 1980): our unpublished data and from Rechnitz zone (KOLLER 1985).

fractionation processes. That is apparent from Nb-Zr diagrams (*Fig. 7*) which indicate close similarities with the other basalts of the Piedmont ophiolites and with ocean floor basalt in particular N-MORB. This correlation is less evident in the case of Meliata volcanics which are typically similar to the North Atlantic basalts (WOOD *et al.* 1979a) reflecting their higher Nb contents with respect to the volcanic rocks of the Piedmont and Rechnitz zones. Estimation of the degree of partial melting on a Cr-Y diagram (*Fig. 8*) shows that Rechnitz zone have higher Cr values comparing with the other two. Most volcanic rocks of the Piedmont and Rechnitz and a few samples from Meliata are situated on the subvertical line marking the fractionation pass for typical oceanic tholeiites (PEARCE 1983; PEARCE and NORRY 1979) which starts at about 17% partial melting. A few points of both Rechnitz and Meliata volcanics are fallen into the island arc field, that can be probably ascribed to the metasomatism (in Rechnitz) and partly from spilitic processes and/or secondary alteration (in the Meliata zone).

Patterns of incompatible trace and minor elements normalized spider-diagrams (after THOMPSON *et al.* 1984 and PEARCE 1983; *Fig. 4* and *9*) are aside from the already discussed similarities of REE patterns (*Fig. 3*), nearly similar. In all the three regions, these volcanics show slight enrichment of the elements situated

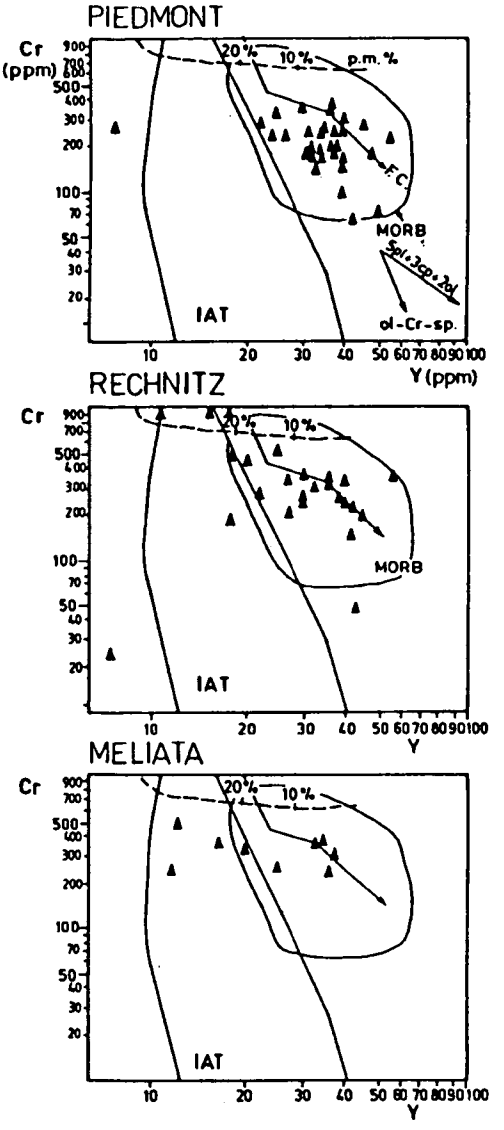

Fig. 8. Cr-Y diagrams (in ppm) of the basaltic rocks for the three regions to discriminate oceanic ridge basalts (MORB) from island are tholeiites (IAT). The horizontal dashed line shows the partial melting (p.m.) path with the amount of mantle melting indicated in %. The subvertical line marks the fractional crystallization (f.c.) path for typical oceanic tholeiite (after PEARCE 1983; PEARCE and NORRY 1979), which starts at about 17% melting. Closed system fractionation segments are subvertical (pl-sp), open system crystallization segments subhorizontal (plag+ol+cpx, HÖCK and MILLER 1987). Sources of data as *Fig. 7.*

on the left side of the diagram (mobile elements: series Sr-Ta of *Fig. 9*), toward plume-type MORB-composition a tendency which has been called "transitional" (T-type) by SUN *et al.* (1979) as well as WOOD *et al.* (1979a, b); LE ROEX *et al.*

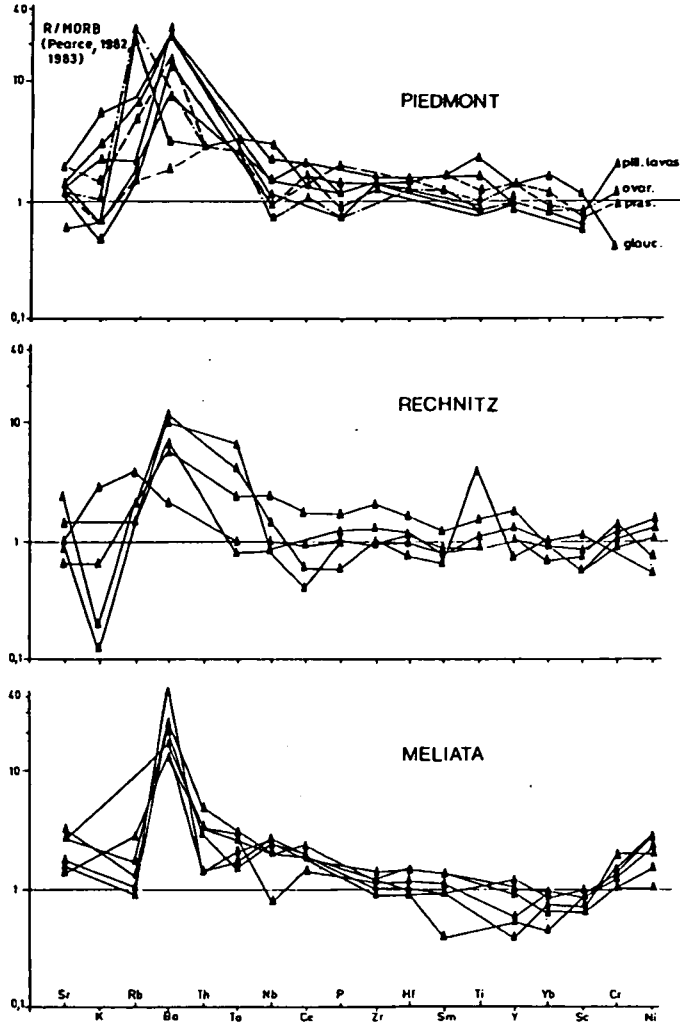

Fig. 9. Comparison of the common trace element contents (as ppm ratios) of basaltic rocks for the three regions with the "normal-MORB" composition of PEARCE (1983). Sources of data as *Fig. 7.*

(1983). In particular, these show closer similarities with N- and T-type (Piedmont zone) E- and T-type (Meliata zone) and with N-type MORB composition (Rechnitz zone). The same tendency was found by POGNANTE *et al.* (1986) and PFEIFER *et al.* (1989) from samples from Piedmont zone and by KOLLER (1985) for the Rechnitz zone. The three regions are enriched in Ba element, moreover, some samples of the Piedmont zone show an enrichment in Rb and K (*Fig. 9*), which can be attributed to metasomatism.

After a review of the ophiolites within the Alpine and Carpathian belts, Penninic and Vardar ocean in the Western Tethyan basin is to be compared.

(1) In the Tethyan basin, Jurassic Piedmont-Ligurian ocean in the Western Alps (including Arc valley, Monviso, Montgenévre and Roche Noire ophiolites) as well as the East one (including Rechnitz ophiolite) opened to the south and to the westward respectively. On the other hand, Triassic Vardar ocean (including Meliata ophiolite) opened to the south. However, both the Penninic and Vardar oceans (in the South Apuseni Mts. and East Carpathians) are overlapped each other (KÁZMÉR and KOVÁCS 1989).

(2) The Western Alps ophiolite, in the Piedmont zone, is composed of polyphase metamorphic dismembered ophiolite bodies and their most Liassic to Cretaceous sedimentary cover (mostly calc-schists). In the easternmost part of the Alps (i.e. the Rechnitz zone) it consists of a less metamorphosed sequence and Mesozoic metasediments. On the other hand, the Meliata zone of the Western Carpathians is composed of fragments of dismembered ophiolite sequence and its Middle Triassic deep-sea sediments (HOVORKA 1985; KUBOVICS *et al.* 1989) and Jurassic shale (KOVÁCS 1990 pers. comm.).

(3) All ophiolite suites, apart from some oceanic alteration, have more or less undergone a complex history of Alpine regional metamorphism with variable grade (eclogite to greenschist facies in the Arc valley and Monviso; glaucophane to greenschist facies in the Roche Noire, Rechnitz and Meliata and greenschist facies to prehnite-pumpellyite facies in the Montgenévre and Meliata).

(4) All units of the Monviso and Eastern Arc valley in the Eastern Piedmont ophiolite share a similar metamorphic history consisting of four stages: (1) oceanic event; (2) eclogitic stage; (3) blueschist stage and (4) greenschist stage. The oceanic metamorphism produced only local development of brown and green hornblende. The eclogitic stage was very pervasived and produced mineral assemblages composed of Na-pyroxene, garnet, epidote and rutile. The blueschist stage was characterized by the widespread development of Na-amphibole (mostly at the expense of Na-pyroxene), albite and locally garnet and phengite. The greenschist event produced irregular development of chlorite, albite, Fe-epidote and locally stilpnomelane. The Rechnitz, Roche Noire and Meliata units are mostly characterized by mineral assemblages of the blueschist to greenschist facies conditions. The best preserved, weakly metamorphosed units are observed in Montgenévre and Meliata ophiolites. These are characterized by prehnite-pumpellyite to greenschist facies mineral assemblages including albite, chlorite, epidote, prehnite, pumpellyite and iron oxide.

(5) Consequently, the grade of metamorphism is perhaps decreased from the Western Alps (Piedmont zone) through Eastern Alps (Rechnitz zone) to Inner Carpathians (Meliata zone).

(6) The amount of serpentinites, compared with the extrusives, are high in the Alpine and Carpathian belt. Serpentinite with harzburgitic composition are more dominant in the Eastern Alps and Inner Carpathians, while those with mostly lherzolitic and subordinately harzburgitic composition are more frequently in the Western Alps. The ultramafics of both the Alp and the Carpathian belt are mostly less abundant in REE as compared to chondritic values with more distinct light REE depletions. They are depleted in Ti (Piedmont zone) and V (Meliata zone) and enriched in Ti and Fe (Rechnitz zone) with respect to the estimated primitive mantle (JAGOUTZ 1979). Ophicarbonate is common in the Alpine belt.

(7) The magmatic (plutonic and volcanic) rocks in the Alps (i.e. Western and Eastern Alps) are characterized by wider compositional range of major, trace and rare earth elements as compared to that from the Carpathians (Meliata zone). The intensity of magmatism is decreased from the Western Alps through Eastern Alps to the Western Carpathians.

(8) The amount of intrusives compared to the extrusives decreased from Piedmont through Rechnitz to Meliata zone.

(9) Proper sheeted dyke complexes are missing in the Alp-Carpathian system.

(10) Most extrusive rocks from Eastern and Western Alps display compositions comparable to "normal" and "transitional" (N-/T-) MORB, perhaps indicating origin from heterogeneous sources by different degree of crystal-fractionation and/or lower degree of partial melting. Conversely, the Meliata basalts which strongly enriched in light and middle REE, show "enriched" and "transitional" (E-/T-) MORB character, and should have been derived by partial melting of differently (or anomalously) enriched sources, probably accompained by little fractionation. These "enriched" basalts are also recorded in the Piedmont zone (BECCALUVA *et al.* 1984b; POGNANTE *et al.* 1986) supporting evidences for heterogeneities in the local sources, as observed in the North Atlantic by WOOD *et al.* (1979).

(11) Ophiolites of the Alpine and Carpathian belt have mostly tholeiitic character with high Ti and Ba contents. They are characterized by a progressive enrichment in the incompatible and rare earth elements starting from ultramafics through intrusives to extrusives and plagiogranites. However, these elements follow a typical magmatic trend, confirming that the original bulk rock-trace elements composition can be preserved even during the different stages of metamorphism.

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FLUORITE GRANITES FROM SOUTHWESTERN SINAI, EGYPT WITH PARTICULAR REFERENCE TO THE A-TYPE

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ABSTRACT

The petrographical and geochemical characteristics of some fluorite granites form Wadi bábá area in the southwestern Sinai of Egypt are presented and compared with the A-type suites. The studied granites belong to the phase-III of Egyptian Younger Granites and are associated, in space and time, with a comagnetic phase-II.

Petrochemically, the fluorite granites possess a peralkaline and sodic-potassic character, with high silica, alkalies and normative acmite. They are magmatic origin, typically A-type granites and are formed through partial melting of the upper mantle to lower crustal materials followed by crystal fractionation of a fluorine-rich extracted magma in a probable extensional environment. They are, moreover, crystallized under moderate temperature and pressure (1-6 Kb.) and are enplaced as moderate to shallow intrusives in the continental crust.

INTRODUCTION

Granite rocks may be subdivided into those generated during evolution of fold belts (orogenic) and those associated with uplift and major strike-slipe faulting (anorogenic). Anorogenic A-type granites, first definited by LOISELLE and WONES (1979) and COLLINS *et al.* (1982) and discussed by HARRIS and MARRINER (1980) and WHALEN *et al.* (1987). However, A-type suites may represent the final plutonic event in both orogenic belts and rift-related anorogenic magmatism or shield areas, emplacing in a wide range age from Proterozoic to Recent (WHALEN *et al.* 1987). A-type granites are, moreover, derived from recycled, dehydrated continental crust which extracted from igneous protoliths (I-type) or formed from highly contrasting mafic-sialic association of two mixing magmas (WHALEN and CURRIE 1984).

A-Type granites have been discussed in the Arabian Shield by JACKSON *et al.* (1984) and other. In the Egyptian Shield, these granites have been treated by several authors. A-type granites are considerated as latest phase intrusives in the main younger granites (RAGAB 1987) or as post-kinematic shallow cauldron complexes corresponding to G-3 granites (EL-GABY *et al.* 1988). These granite plutons, in Sinai, are grouped together with alkaline monzonite and rhyolite as Katherina Province (BENTOR 1985). In the southwestern district of Sinai, under study, the younger granites were emplaced during late Pan-African event between 609 and 568 Ma and are chemically corresponded the phase II and III (ABDEL-KARIM and ÁRVA-SÓS in press) of the Egyptian Younger Granites. EL-AREF *et al.* (1988) described some plutons of the younger granites of the SW-Sinai as G-2 granite. Moreover, a good relationship between fluorine and latest pulses of the

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Egyptian younger granites have been recently discussed by BENTOR (1986), and RAGAB (1987).

GEOLOGIC SETTING

Younger granites of Wadi Bábá area (including fluorite granites) consist of several large irregular shape complexes of batholithic dimensions, dissected by Wadies: Samra, Bábá, Lahian, Sahu, Budra, Shellal, Igna, Sidri and Atima (*Fig. 1*). The plutons exhibit extremely sharp contact with country rocks including schists, gneisses, migmatites, metagabbro-diorite and mafic dyke-swarms and, in places, by post Cambrian sediments. The granite plutons are collectively composed of dominant monzo- and syenogranite and minor alkali feldspar granite and quartz syenite. The granite rocks are medium to coarse grained, occasionally porphyritic and granophyric and show grayish pink to red colour. Field relation and chemistry indicate that these plutons are corresponded to the phase II and III of the Egyptian Younger Granites. Phase-II is usually crosscut by a number of felsic dykes compared with phase-III. The fluorite granites most probably equivalent to phase-III which merge imperceptibly into a comagmatic phase-II (ABDEL-KARIM in press) with rather indefinite contacts.

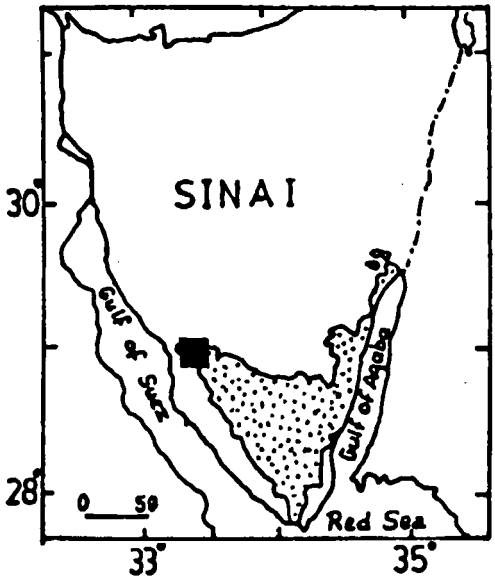

Fig. 1. Location map of the examined fluorite granites.

PETROGRAPHY

The modal analyses of seven samples, collected from examined plutons are shown in Table 1 and plotted in the classification diagram of STRECKELSEN (1978). It is evident that most of the studied rocks are alkali feldspar granite composition.

TABLE 1.

Modal analyses of fluorite granites from Wadi Bábá area, SW-Sinai.

Rock-type	Sample No.	Alk. feld.	Q.	Plag.	Biot.	Amph.	Fluo.	Acc.
Medium to coarse grained granites	1	52.7	32.5	14.3	0.5	—	0.5	0.3
	2	48.8	42.0	6.5	0.1	1.6	—	1.0
	3	50.5	41.1	4.3	1.5	0.4	0.8	0.5
Porphyritic granites	4	42.2	43.3	10.6	2.5	—	1.5	1.2
	5	45.3	38.7	11.0	1.3	1.5	0.7	2.1
Granophyric granites	6	41.3	41.6	14.4	2.7	0.6	0.2	1.7
	7	44	38.3	12.8	1.5	0.5	0.6	1.3

Petrographically, the examined rocks are mostly perthite granites (hypersolvus). They are medium to coarse grained hypodiomorphic-granular texture. Granophyric and porphyritic textures are common, too. The perthite granites consist mainly of perthitic alkali feldspar and quartz with minor plagioclase, biotite, alkali amphibole relict and fluorite. Iron oxide, topaz titanite, monazite and apatite are the main accessory phase.

The alkali feldspar represents patchy, flamy and stringly shaped orthoclase perthite with minor microline perthite. Biotite occurs as brown interstitial flakes. Alkali amphibole forms bluish green anhedral crystal relics. Fluorite occurs as yellowish, highly relief and isotropic crystal aggregates, mostly associated with biotite. In the porphyritic varieties, megacrysts of perthitic alkali feldspar (25.5—26.6 %), quartz (8.5—9.4 %) and plagioclase (4.5—4.8 %) are embedded in a microgranitic groundmass consisting of quartz (30.2—33.8 %), alkali feldspar (16.6—18.7 %), plagioclase (5.8—6.5 %), biotite (2.5—1.3 %), alkali amphibole (0.5—0.8 %), fluorite (1.1—1.4 %) and iron oxide. Granophyric granites consist of small amount of megacrysts (10—13.6 %) of alkali feldspar, quartz and plagioclase embedded in a fine-grained graphic and granophyric groundmass (85.5—90 %) composed of the same minerals of porphyritic one.

PETROCHEMISTRY

The results of 9 newly analyzed samples from fluorite granites of the SW-Sinai are compared with the average values of the A-type granites of COLLINS *et al.* (1982) (Table 2). The analyses of major elements were carried out in the laboratories of the Hungarian Geological Survey and Eötvös University in Budapest, using standard wet chemical techniques.

Major elements: The studied granite rocks contain, on average, 76 % SiO_2, 11 % Al_2O_3, 1.2 % FeO_t, 0.1 % MgO and 0.9% CaO (Table 2). Consequently, the average chemical analyses of the plutons are well comparable with that of the A-type granites of COLLINS *et al.* (1982), with slightly higher CaO, MgO and Na_2O contents. The compositional ranges of the main normative minerals are: Q (29—37 %), Or (26—35 %), Ab (18—36 %) with minor Ac (0.7—3 %) and Ns (0.2—2 %). THORNTON and TUTTLE DI of the rocks ranges from 88 to 96.

The examined granites are plotted on the alkalinity variation diagram of WRIGHT (1966) (*Fig. 2*). It is clear that the present granit rocks exhibit a strong alkaline to peralkaline nature which is reflected by the occurrence of normative acmite and Na-metasilicate. Morover, plot of these rocks in the Ca + Mg + Fe_t -

Na + K - Al diagram after BOWDEN and TURNER, (1974) (*Fig. 3*) suggests again the peralkaline nature of these granites. However, the chemical analyses (Table 2) confirm the slightly peraluminous to metaluminous affinity of the studied granites (A/CNK ratio: 0.91—1.23).

TABLE 2.

Major element compositions of fluorite granites from Wadi Bábá area, SW-Sinai

Oxides	1	2	3	4	5	6	7	8	9	Aver.	A-type*
SiO_2	78.40	78.00	76.75	76.60	75.90	76.00	74.80	75.08	74.24	76.20	77.21
TiO_2	0.10	0.10	0.10	0.10	0.10	0.10	0.17	0.10	0.10	0.11	0.19
Al_2O_3	11.00	11.43	10.13	10.75	12.10	11.70	11.90	10.75	10.38	11.13	11.79
Fe_2O_3	1.06	0.25	0.78	0.37	0.82	0.95	0.63	0.87	0.88	0.73	0.36
FeO	0.33	0.21	0.81	0.82	0.27	0.26	0.66	0.65	0.57	0.51	0.85
MnO	0.01	0.01	0.03	0.02	0.02	0.01	0.04	0.04	0.03	0.02	0.03
CaO	0.25	0.49	0.73	0.88	0.67	0.87	1.45	0.75	1.90	0.89	0.39
MgO	0.10	0.13	0.15	0.12	0.10	0.06	0.10	0.10	0.13	0.11	0.04
Na_2O	3.96	4.40	3.56	3.82	4.67	4.45	4.85	3.84	4.10	4.18	3.08
K_2O	5.33	4.37	6.03	5.62	4.72	4.99	4.50	5.52	5.35	5.16	5.00
$+H_2O$	0.00	0.00	0.00	0.00	0.00	0.42	0.64	0.00	0.00	0.12	—
$-H_2O$	0.13	0.16	0.31	0.12	0.18	0.08	0.23	0.17	0.27	0.18	—
P_2O_5	0.01	0.02	0.01	0.01	0.01	0.01	0.02	0.01	0.01	0.01	0.02
Sum	100.68	99.57	99.39	99.23	99.56	99.90	99.99	97.86	97.96	99.35	
					C I P W norms						
Q	37.15	36.28	37.40	35.63	30.93	31.95	29.56	34.09	32.97		
Or	31.50	25.82	35.63	33.21	27.89	29.49	26.59	32.62	31.62		
Ab	26.91	34.46	18.54	24.01	35.96	32.40	36.16	24.56	23.61		
Ac	3.07	0.72	2.26	1.07	2.37	2.75	0.00	2.62	2.55		
Ns	0.73	0.45	2.10	1.65	0.20	0.50	0.00	1.18	1.91		
Wo	0.20	0.55	0.41	0.50	1.07	1.06	2.23	0.94	3.23		
Cpx	0.54	0.77	2.19	2.65	0.54	0.32	1.45	1.18	1.09		
Mt	0.81	0.36	1.13	0.54	0.64	0.58	0.91	1.26	1.28		
Il	0.19	0.19	0.19	0.19	0.19	0.19	0.32	0.19	0.19		
Hm	0.50	0.00	0.00	0.00	0.38	0.55	0.00	0.00	0.00		
Ap	0.02	0.05	0.02	0.02	0.02	0.02	0.05	0.02	0.02		

*A-type: average of 8 analyses of A-type granites (COLLINS *et al.* 1982).

The FAM diagram (*Fig. 4*) for the studied rocks reveals that they are enriched in alkalies and strongly depleted in magnesium and iron, these features probably result from extrems fractionation of the present rocks. Moreover, the parallel distribution of the analyzed samples close to the FA side of the diagram perhaps indicates an extensional regime PETRO *et al.* 1979) for the studied granites.

Classification: *Fig. 5* illustrated the chemical classification of the examined granites which is represented on the basis of their normative An, Ab and Or (BARKER 1979). In this diagram, all the analyzed samples fall within the granites field. Furthermore, the studied plutons lie near the field of alkali granites showing

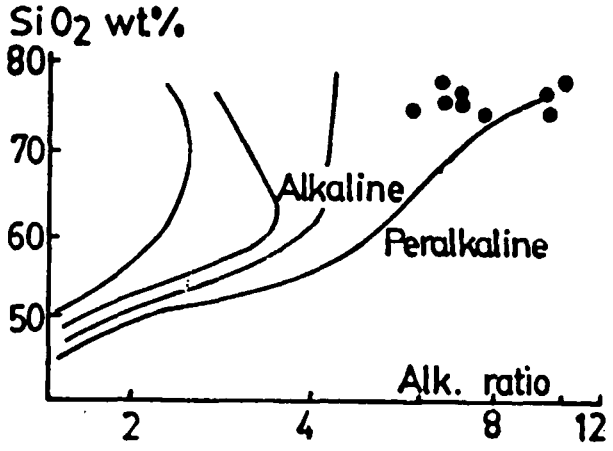

Fig. 2. SiO_2 vs. alkalinity ratio variation diagram (WRIGTH 1969) for the examined fluorite granites.

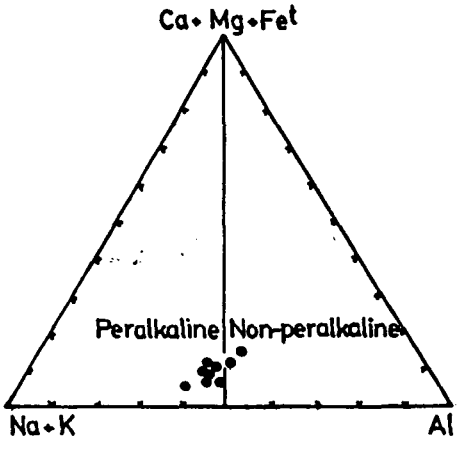

Fig. 3. Na+K - Ca+Mg+Fe_t - Al ternary diagram (BOWDEN and TURNER 1974) for the examined fluorite granites.

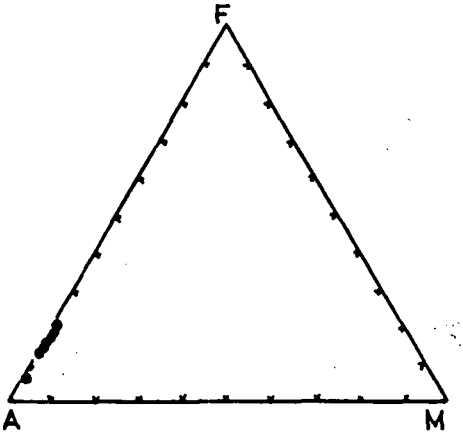

Fig. 4. APM ternary diagram for the examined fluorite granites.

their loss of MgO with increase SiO_2, probably during extreme fractionation. That is appeared on the SiO_2-log10 (K_2O/MgO) diagram proposed by ROGERS and GREENBERG (1981) (*Fig. 6*). Consequently, these rocks are corresponded to the third phase younger granites (HEIKAL *et al.* in press) on the same diagram.

PETROGENESIS

The normative An, Ab and Or proportion of the granitic rocks are plotted in *Fig. 7.* It is clear from the figure that the studied samples plotted on midway along the Ab-Or side indicating their transitional (e.g. sodic-potassic) character. Also, all the samples lie in the orthoclase field mostly near the isobaric univariant curve

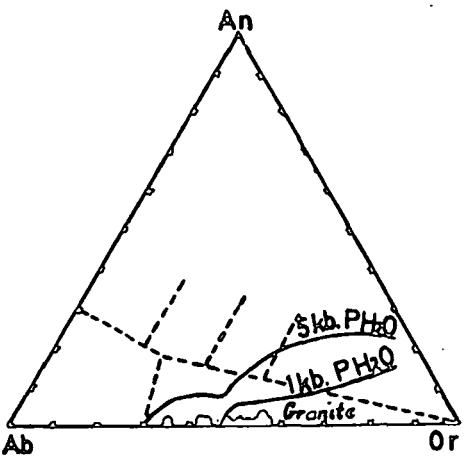

Fig. 5. Ab - An - Or ternary diagram (BARKER 1979) for studied granites

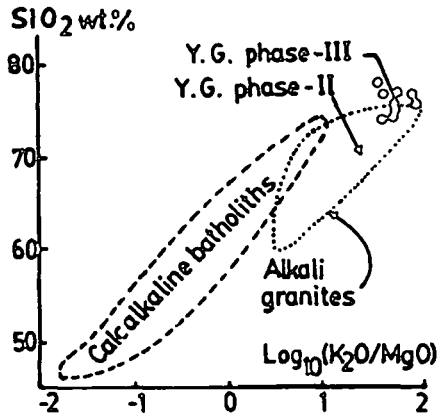

Fig. 6. SiO_2 vs. Log 10 (K_2O/MgO) variation diagram (ROGERS and GREENBERG 1981) for the studied fluorite granites. Trends of Phase II and III Younger Granites after HEIKAL *et al.* (in press)

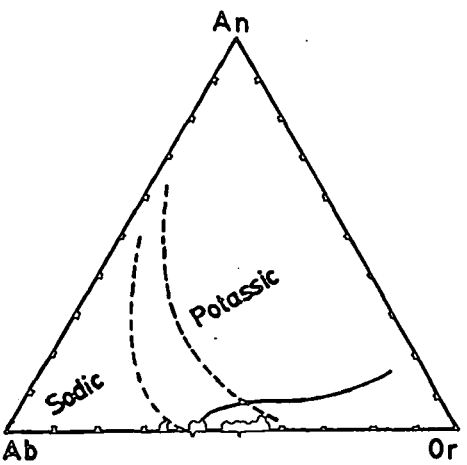

Fig. 7. Ab - An - Or ternary diagram for the studied fluorite granite. Sodic and potassic zones after IRVINE and BARAGER (1971). Solid line represents the two feldspar binary curve for quartz saturated feldspar system at 1 Kb. water-vapour pressure (after JAMES and HAMILTON 1969).

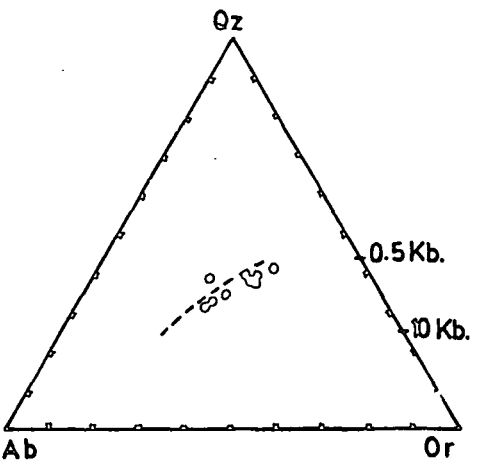

Fig. 8. Ab - Qz - Or ternary diagram for the studied fluorite granites. Dashed line represents the variation in the minimum melting points of the water-vapour pressures from 0.5 to 10 Kb. (after TUTTLE and BOWEN 1958).

indicating that the crystall-liquid equilibrium was the dominant mechanism involved in the genesis of these granites (JAMES and HAMILTON 1969).

The normative Q, Ab and Or proportion of the examined rocks are plotted in *Fig. 8.* and the results are compared with the experimental data of TUTTLE and BOWEN (1958). It is observed from the figure that the composition points of the studied rocks fall near the minimum melting point at low to moderate pressure

(1—6 Kb) in the Ab-Or-Q-H_2O system of TUTTLE and BOWEN (1958). Such a pressure, in turn suggests that these rocks were emplaced at shallow depth in the crust. Furthermore, most samples scatter slightly toward the Or-Q side.

The conclusion of the foregoing discussion becomes clear, when comparing the mineralogical and geochemical characteristics of the investigated fluorite granites (e.g. third phase younger granites) with those for typical A-type granites (LOISELLE and WONES 1979; COLLINS *et al.* 1982; JACKSON *et al.* 1984; WHALEN *et al.* 1987) (Table 3). It is evident that the studied fluorite granites are most probably analogous to A-type granite plutons.

TABLE 3.

Mineralogical and chemical properties of typical A-type granites (LOISELLE and WONES 1979; COLLINS et al. 1982; JACKSON et al. 1984; WHALEN et al. 1987) compared with those of some fluorite granite rocks from SW-Sinai, Egypt.

A-type characteristics	Studied fluorite granites
- Compositional range: from granite to adamellite. - Felsic min.: perthitic alkali feldspar very common. - Mafic min.: biotite late and interstitial. - Acces. min.: fluorite common, +/- arfedsonite, riebeckite. - Micrographic intergrowths very common.	- Alkali feldspar granite. - Perthitic orthoclase (+/- microcline) common. - Interstitial biotite common. - Fluorite common, +/- alkali amphibole, monazite and topaz. - Micrographic intergrowths common
- High SiO_2, Na_2O and K_2O contents. - Low Al_2O_3, CaO and MgO contents. - Agpaitic index: <1 - Na_2O/K_2O: 1 - Mostly near peralkaline. - Contain normative acmite. - High Thornton and Tuttle DI (88-95).	- High SiO_2 (77%), Na_2O (3%), K_2O (5%), - Low Al_2O_3 (11.8%), CaO (0.4%), MgO (0.04%) - Agpaitic index: 0.97—1.36 - Na_2O/K_2O: 0.6—1.0 - Peralkaline. - Normative acmite: 0.7—3% - High DI (88—96)
- Usually associated in space and time with alkaline complexes (biot. granite, peralk. granite and qz. syenite)	- Commonly associated in space and time with the alkaline younger granites phase-II.

Several magmatic processes have been suggested to be involved in the origin of the alkali granites including partial melting, crystal fractionation, metasomatism, liquid immiscibility, halide complexing, and thermo-gravitational diffusion (CLEMENS *et al.* 1986). However, for A-type granites, direct partial melting of middle of lower crustal rocks, possibly followed by crystal fractionation in the presence of fluorine-rich volatiles to form the more evolved varieties, is perhaps the most likely model (COLLINS *et al.* 1982; WHALEN *et al.* 1987).

The A-type granites of the Afro-Arabian Shield probably were generated by partial melting, under high temperatures, vapour-absent conditions of a crust from which granodiorite melts had previously been extracted (DRYSDALL *et al.* 1984). The partial melting processes of the deep crust of the Afro-Arabian Shield would be initiated either by the late Precambrian accretion of arc systems which resulted in a thickening of the juvenile crust (GREENBERG 1981) or by a combination of three factors: rising temperatures, dimension pressures and depression of melting points by the presence of the fluorine among the volatiles (BENTOR 1985). Since the studied granites are hypersolvus, peralkaline, rich in fluorite, high in silica and alkalics and are indistinguishable from the A-type granites, these granites have a geochemical indicative of their generation by partial melting of the upper mantle

to lower crustal rocks followed by crystal fractionation of fluorine-rich extracted magma. Trace and REE data are needed to test this hypothesis.

CONCLUSION

The mode of occurrence, petrographic and geochemical characteristics of some fluorite granites from Wadi Bábá area in the SW-Sinai have been presented and compared with the A-Type granites (LOISELLE and WONES 1979; COLLINS *et al.* 1982). The examined granites represent the latest pulses of the Egyptian Younger Granites as well as belong to the phase-III which, in the field, merge imperceptibly into a comagmatic phase-II with a rather indefinite contact. These rocks are medium to coarse grained; granophyric and porphyric varieties are common, too. They are hypersolvus-type consisting of perthitic alkali feldspar, quartz and plagioclase with minor biotite, alkali amphibole and fluorite.

Petrochemically, the fluorite granites are sodic-potassic character, enriched in silica, alkalies and normative acmite and Na-metasilicate. They show strong alkaline to peralkaline nature (A/CNK: 0.91—1.23) with an extremely crystal fractionation (DI: 88—96). The studied granites are typically A-type granites, most probably derived by partial melting of upper mantle to lower crustal rocks followed by crystal fractionation of an extacted magma enriched in fluorine. Furthermore, they were crystallized under moderate water vapour pressure 1—6 Kb. (TUTTLE and BOWEN 1958) and were emplaced at relatively moderate to shallow depth in a continental crust of an extentional regime.

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PETROGRAPHICAL CHARACTERISTICS OF DITRÓ (OROTVA) HORNBLENDITES, EASTERN CARPATHIANS TRANSYLVANIA (ROMANIA): A PRELIMINARY DESCRIPTION

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ABSTRACT

During the 150 years the syenite massif of Ditró has been examined by excellent mineralogists and petrographers like HERBICH (1872); KOCH (1880); BERWERTH (1905); REINHARDT (1911); MAURITZ (1909-25); JANOVICI (1932, 1934, 1970); STRECKEISEN (1931-74); CODARCEA *et al.* (1957); ANASTASIU and CONSTANTINESCU (1979-81); JAKAB (1976, 1982, 1986).

Magmatic mass of the massif penetrates the metamorphic rocks of the Tölgyes Series and the Rebra-Barnar Series and the Bretila Series (i.e. Rarau Gneisses and Hagymás Granitoides). Certain parts of the massif are covered by Pliocene limnic sediments (JEKELIUS 1923), Late Pliocene-Early Quaternary gravels (STRECKEISEN 1952) and agglomerates and tuffs as well as the lava layer of the Kelemen-Hargita volcanic range (JANOVICI 1934; SAVUL and KRÄUTNER 1937; RADULESCU *et al.* 1973) (*Fig. 1*).

According to our present knowledge the age of the Ditró syenite massif is between 150—160 and 120—130 My. (BAGDASARIAN 1972; STRECKEISEN and HUNZIKER 1974; MINZATU 1982 and JAKAB 1982).

On the basis of their spatial setting and structure and their petrochemistry as well as relationships to each other, the rocks of the Ditró syenite massif can be divided into several groups. A wide range of variety of the rocks located into a fairly small area suggests different intrusions. Their alkaline character can be attributed to the results of several alkaline-metasomatic phases which touched not only the magmatic-metamorphic complexes found within the massif, but the surrounding rocks, too.

Outcropping of the hornblendites in the basin of the Orotva creek (northern part of the massif) takes a distinctive importance among these groups of rocks. According to the recent — but not proved — theory (JAKAB 1986) the hornblendites are originated by Fe and Mg metasomatism of earlier crystalline rocks which elements derived from the adjacent alkaline environment. It is obvious that further petrological investigations need for the correct genetical estimation of the Orotva creek hornblendites.

INTRODUCTION

The Ditró syenite massif is situated on the northern-northeastern side of the Gyergyó Basin on the southwestern edge of the Gyergyó Alps in the middle part of the Eastern Carpathians (*Fig. 1*). In its northern part (between the Csibi-Jakab creek and Török creek in the Orotva valley) groups of rocks outcrop and they form a succession from the west to the east according to the quantitative decrease of the non-ferrous rock-forming minerals. The Orotva Complex can be subdivided into five groups (ANASTASIU *et al.* 1979. *Fig. 2*) as follows: (1) hornblendites, (2) diorites, (3) monzonites and syenites, (4) foiditic rocks, (5) granodioritic rocks. The aim of this paper is to give a mineralogical and petrographic presentation of

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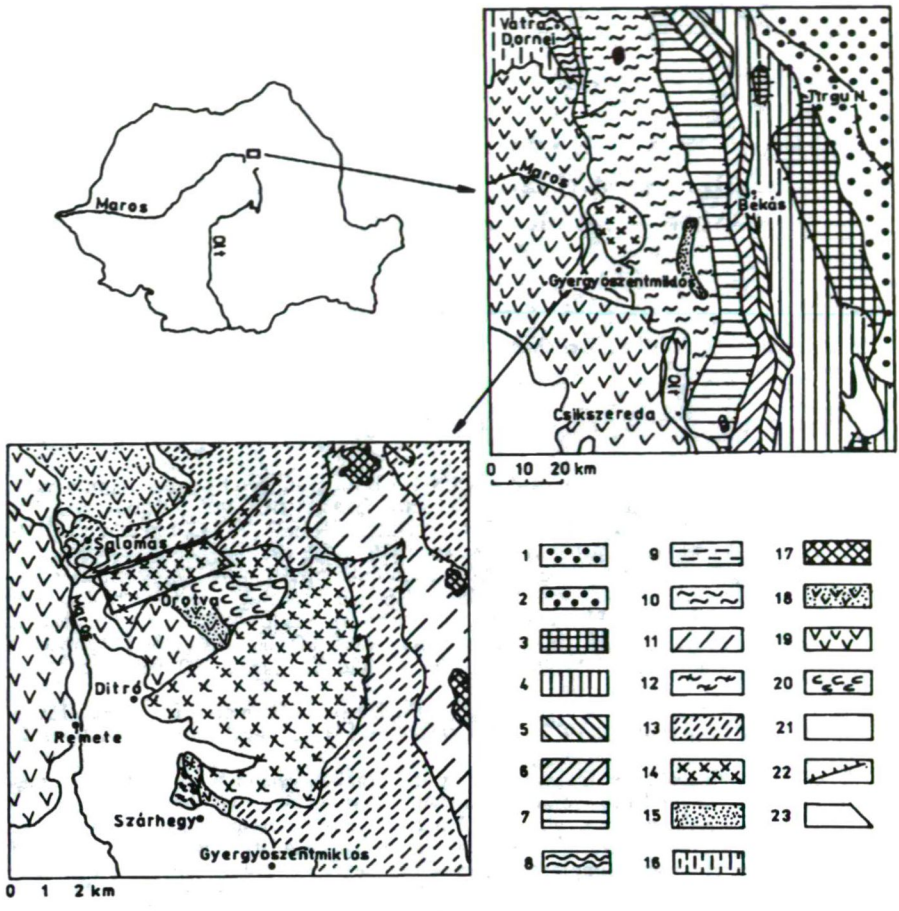

Fig. 1. Site of the Ditró syenite massif (Romania) and its connections with the structural units of the Eastern Carpathians (SANDULESCU *et al.* 1981) and geological of Ditró and its environs (IGC).

Legend. Fore-deep: (1) non-folded Moldavids, (2) Sub-Carpathian Nappe, (3) Marginal Nappe, (4) Tarkő Nappe, (5) Audia Nappe, (6) Convolute Flysch Nappe. Outer Dacids: (7) Csalhó Nappe. Inner Dacids (Bukovinian Nappe System): (8)Infra-Bukavinian Nappe, (11) Bretila Serie, (12) Rebre-Barnari Serie, (13) Tölgyes Serie, (14) Ditró syenite massif, (15) Transylvanian Nappes. Post-tectonic cover: (16) Lápos Autochtonous, (17) Triassic sediments, (18) Neogene–Qaternary rocks in general, (19) Neogene and Quaternary volcanic rocks, (20) Pliocene sediments, (21) Neogene basins, (22) Overthrust zone, (23) Investigated area.

the group of hornblendites, in detail. In the view of this paper the hornblendite group consists of hornblendite rocks containing more than 90 % hornblende minerals as well as their pegmatoidic derivatives (although their mafic content sometimes is smaller than 50 %).

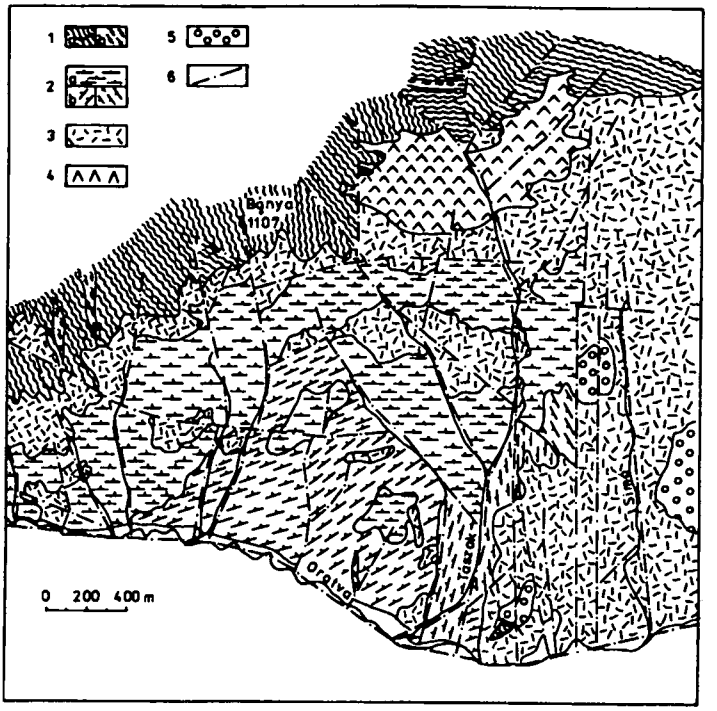

Fig. 2. Geological map of the northern part of the Ditró syenite massif (ZÓLYA, IONESCU, PÁL- MOLNÁR 1988).

Legend. (1-a) Crystalline rocks, (1-b) Black quartzites, (2) Group of hornblendites and diorites: (2-a) non-oriented hornblendites, (2-b) oriented hornblendites, (3) Group of monzonites and syenites, (4) Group of foidites, (5) Volcanogene-sedimentary rocks, (6) Fault.

PETROGRAPHIC CLASSIFICATION OF DITRÓ-OROTVA HORNBLENDITES

Based on their texture the hornblendites are divided into two large groups (CODARCEA *et al.* 1975): (I.) rocks with textural ordering, (II.) rocks without textural ordering. According to their mineral composition and structural characteristics these groups can be subdivided into the types as follows:

I. Hornblendites with textural ordering
   1. lineated hornblenbites with microcrystalline structure and abundant sphene content,
   2. phanerocrystalline hornblendites with sphene and biotite,
   3. sphene- and biotite-free phanerocrystalline hornblendites.

II. Hornblendites without textural ordering
   1. sphene- and biotite-free microcrystalline hornblendites,
   2. sphene- and biotite-bearing microcrystalline hornblendites,
   3. sphene-bearing phanerocrystalline hornblendites,
   4. phanerocrystalline hornblendites with biotite,
   5. sphene- and biotite-bearing phanerocrystalline hornblendites,
   6. olivine-pyroxene hornblendites,
   7. pyroxene hornblendites,
   8. pegmatoide hornblendites.

These rocks are characterized by parallelly embedded hornblende columns and sphene crystals. By the shape of the rock-forming components their structure is regarded to be a subhedral-allotriomorphic one and the degree of their crystallinity shows a decussate crystalline structure (*Fig. 3*). On the basis of the absolute size of the crystals their structure can be regarded phaneritic (the crystals are visible by naked eye) and aphanitic (microcristalline) ones. The lineation of microcrystalline hornblendites is always determined by sphene crystals while in the phanerocrystalline types it is given by amphiboles (and sphenes and biotites if they are present in the rocks). Origin of erdered texture has been explained by several theories: (a) crystallisation of the rocks under high pressure (IANOVIC 1934), (b) metamorphism which retained the schistosity during the anatectic

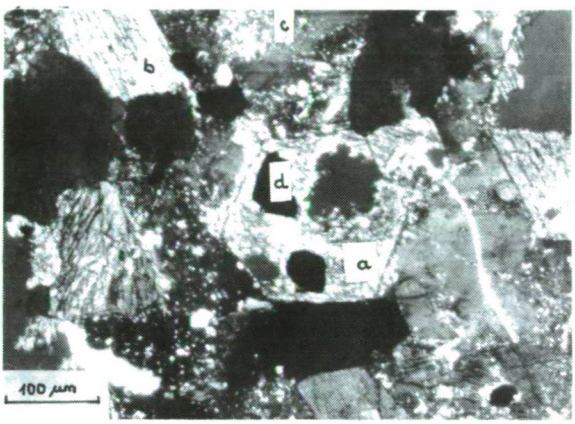

Fig. 3. Hornblendite with textural ordering: (a) hornblende, (b) biotite, (c) plagioclase, (d) magnetite. 28X, +N.

process (CODARCEA *et al.* 1957), (c) metasomatism of crystalline protholith preserving the original schistosity (JAKAB 1986). Modal composition of hornblendites with textural ordering is shown in Table 1.

TABLE 1.

Modal composition of hornblendites with textural ordering

Minerals	volume per cent		
	1	2	3
Hornblende	70—72	80—82	85—86
Plagioclase An 4-58	7—8	2—3	5—6
Orthoclase	1—2	—	—
Biotite+Chlorite	4—5	3—4	—
Titanite	6—7	6—8	—
Apatite	2—3	4—6	4—5
Magnetite	2—3	2—3	3—4

(1) microcrystalline, sphene-bearing hornblendite, (2) phenocrystalline, sphene- and biotite-bearing hornblendite, (3) phenocrystalline, sphene- and biotite-free hornblendite.

Majority of the rocks consist of prismatic hornblende. Dicroism of these minerals shows light brown-redbrown or sometimes light brown subordinately slightly greenish brown colour (*Fig. 4*). It can include apatite by a poikilitic way as well, but in most cases the apatite develops among mafic minerals.

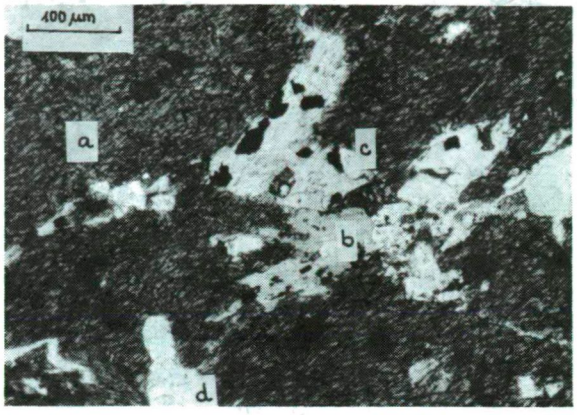

Fig. 4. Hornblendite with textural ordering: (a) hornblende, (b) chloritized biotite, (c) apatite, (d) sphene. 28X, 1N.

Biotite can be found only in closely intergrowth with hornblende and due to the chloritization its dicroism is light brown or green-striped dark brown. It contains sphene and apatite as inclusion. The biotite plates quite often are bended or smashed and always altered. Result of this alteration is a greenish-blue mineral: the pennine. Biotite-pennine changes are always accompained by the occurrence of small sphene crystals which are embedded in the cleavage planes of the biotite or they are in an aggregate form.

Leucocratic minerals are presented by anhedral feldspars. Their polysynthetic twins are sometimes strongly vavy (*Fig. 5*). It is conspicuous that the anorthite content of the plagioclase feldspars of these hornblendites are 4-58 percent (CONSTANTINESCU, ANASTASIU 1977). We think to be a higher anorthite content

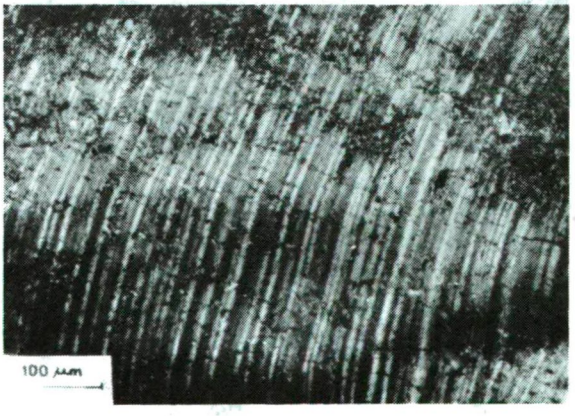

Fig. 5. Hornblendite with textural ordering: plagioclase with strongly vawy polysynthetic twins. 28X, +N.

in them. Some feldspars are sericitized or epidotized or zoisitized (*Fig. 6*). Orthoclase also occurs in very small amount. Sometimes orthoclase-microcline plates dotted by small albite crystals are also recognizable (antiperthite, *Fig. 7*).

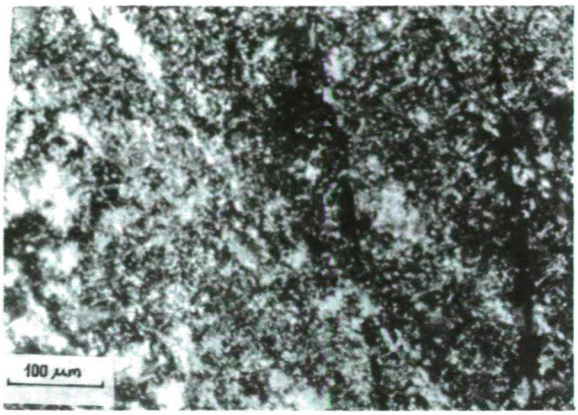

Fig. 6. Hornblendite with textural ordering: sericitized and epidotized feldspars. 28X, +N.

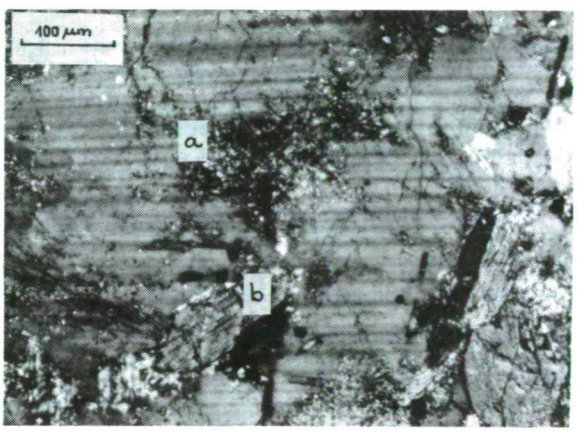

Fig. 7. Hornblendite with textural ordering: (a) antiperthite, (b) biotite. 28X, +N.

Feldspathoides are not characteristic in these rocks.

The sphene always shows two generation origin. The first generation sphene is larger euhedral characteristically rombic sometimes feather-shaped yellowish-brown crystal which is often twinned (*Fig. 8*). The large sphene crystals can include small euhedral amphibole inclusions. Second generation of sphene occurs as small grain in the hornblendites but it mostly appears in altered biotite plates (*Fig. 9*).

The apatite is a quite frequent component of these rocks. They always form euhedral crystals as inclusions found in any other minerals (*Fig. 10*). Magnetite is also not rear and it forms irregular or rounded grains (*Fig. 11*). Based on shape and orientation of these accessory minerals a probable sequence of their crystalli-

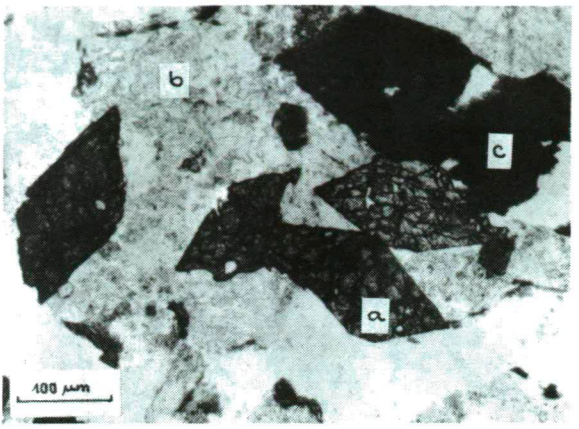

Fig. 8. Hornblendite with textural ordering: (a) sphene, (b) feldspar, (c) biotite. 28X, 1N.

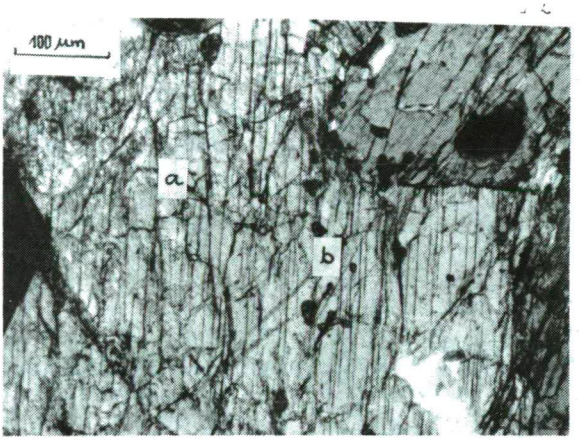

Fig. 9. Hornblendite with textural ordering: (a) biotite, (b) second generation sphene. 28X, 1N.

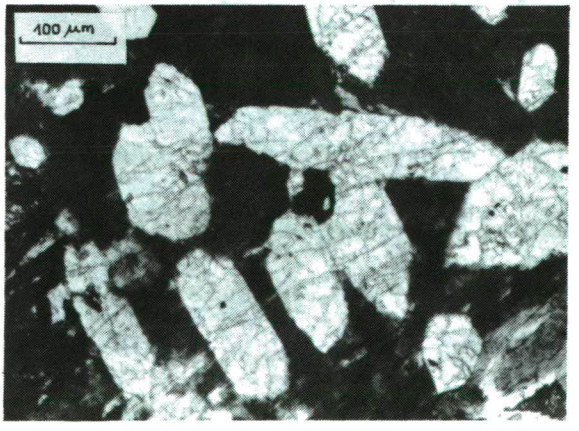

Fig. 10. Hornblendite with textural ordering: apatite. 28X, +N.

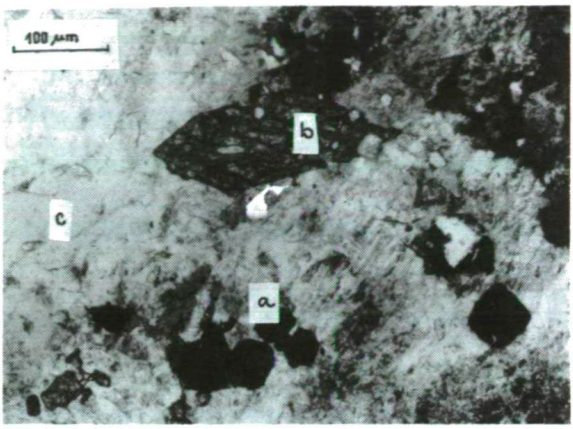

Fig. 11. Hornblendite with textural ordering: (a) magnetite, (b) sphene, (c) feldspar. 28X, 1N.

zation can be supposed, as follows: apatite, magnetite, first generation sphene, hornblende, biotite, second generation sphene and plagioclase.

Epidote veinlets are frequent in the hornblende minerals. Sometimes an epidote excess prevails in the whole rocks forming interstitial fillings or irregular nests between larger crystals. In the chlorite-epidote associations microgranular albite also occur as veinlets cutting calcite crystals. Simultaneous presence of plagioclase (An 86-88) and the albite (CONSTANTINESCU, ANASTASIU 1977) disturb the determination of the order and circumstances of the crystallization. Spatial relationships of these minerals and their relations to other minerals suggest a secondary character of the albite.

Existence of second generation sphene and chloritized biotite as well as frequent epidotization and low anorthite content of the plagioclase, moreover the infrequent presence of the albite and presence of orthoclase and microcline are in close genetical connection with each other. Considering the alkaline character of the whole Ditró Massif, these fhenomena originated from rearrangement of Na, Ca and K content i.e. probably a powerful alkaline metasomatism took place (PÁL-MOLNÁR 1988).

PETROGRAPHIC CHARACTERIZATION OF HORNBLENDITES WITHOUT TEXTURAL ORDERING

It is characteristic that amphibole crystals of the rock-types show a non-oriented arrangement. According to their structure they can be divided into three sub-types: (1) rocks having a phaneretic structure (sometimes in pegmatoidic development characterized occasionally by several cm large amphibole and sphene prisms), (2) rocks consisting of small microgranular crystals, (3) rocks show a porphyric character (larger amphibole and biotite crystals are embedded among smaller rock-forming minerals).

According to the degree of crystallinity they may be docrystalline; in accordance with the shape of crystals they could be subhedral sometimes anhedral; and regarded their size they are phaneritic or microcrystalline rocks.

Varieties listed at the beginning of this sub-chapter No. II, all of them up to the olivine-pyroxene hornblendites have same mineralogical composition to that of hornblendites having oriented texture. The single distinctive criterium is the orientation of the minerals. Their modal composition can be seen in Table 2. Olivine-pyroxene hornblendites and pyroxene hornblendites as well as pegmatoidic hornblendites differ from them both in modal composition and stucture.

TABLE 2.

Modal composition of hornblendites without ordering

Minerals	volume per cent					
	1	2	3	4	5	
Hornblende	87—90	83—84	80—82	68—70	66—68	65—67
Actinolite	—	—	—	—	13—15	—
Biotite	—	2—3	—	6—8	1—2	14—16
Plagioclase An 4-58	5—6	2—3	2—3	8—10	—	6—7
Titanite	—	5—6	8	—	5—6	4—5
Apatite	3—4	2—4	4—6	8—10	7—8	2—3
Chlorite	—	—	—	—	4—6	—
Sericite	—	—	—	3—4	1	2—3
Carbonate	—	—	—	—	2—3	1
Magnetite	2—4	0—3	2—3	—	0—1	0—1

(1) microcrystalliner, sphene- and biotite-free hornblendite, (2) microcrystalline, sphene- and biotite-bearing hornblendite, (3) phanerocrystalline, sphene-bearing hornblendite, (4) phanerocrystalline, biotite-bearing hornblendite, (5) phanerocrystalline, biotite- and sphene-bearing hornblendite.

Olivine-pyroxene hornblendites have a poikilitic structure. Their modal composition is shown in Table 3 and *Fig. 12*. The amphiboles are brown hornblendes shown a considerable dichroism and they are often penetrated by sagenite lattices.

TABLE 3.

Modal composition of olivine-pyroxene hornblendites

Minerals	volume per cent
	6
Hornblende	45—50
Pyroxene	11—38
Olivine	3—21
Biotite	1—2
Plagioclase, An 58	5—25
Apatite	1—2
Opaque, other	1—10

The pyroxenes are Ti-augites which forms equigranular crystal inclusions in the hornblende or biotite grains. Olivine crystals also appear as inclusions of hornblendes forming rounded irregular grains and most frequenly they are transformed into antigorite. The plagioclase (An 58, CODARCEA *et al.* 1957) form euhedral and

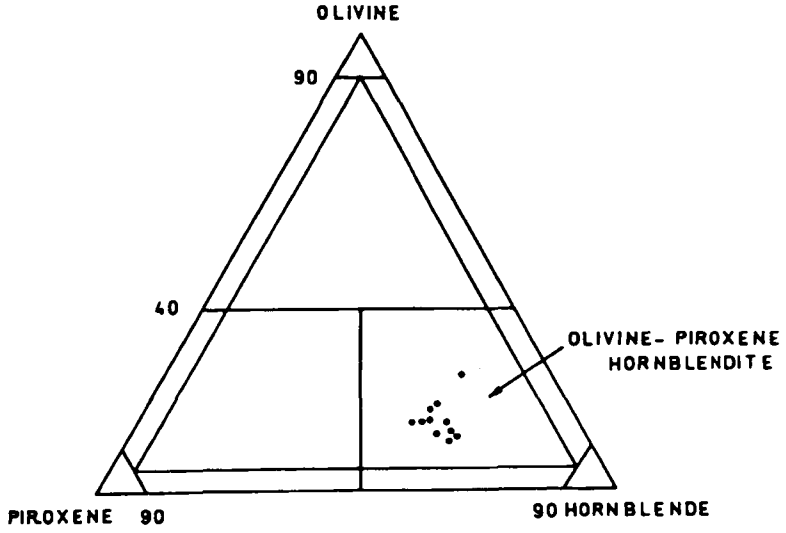

Fig. 12. Classification of ultramafic rocks containing hornblende: olivine-pyroxene hornblendite (after STRECKEISEN 1974).

well developed crystals which are surrounded by hornblende prisms. On the basis of microscopic study the following sequence of crystallization has been found: apatite, magnetite, olivine, augite, plagioclase, hornblende and finally the biotite.

Pyroxene-hornblendite rocks have allotriomorphic structure which is formed by rounded and foiled shape of the rock-forming minerals. The thick prismatic pyroxene crystals surrounded by feldspars show an euhedral habit, only. Modal composition of these rocks are shown in Table 4 and *Fig. 13*. Hornblendes having

TABLE 4.

Modal composition of pyroxene hornblendites

Minerals	volume per cent		
	7		
Hornblende	58—60	54—56	64—66
Actinolite	2—3	4—5	—
Biotite	10—12	30—32	—
Pyroxene	10—12	12—14	20—22
Titanite	4—5	—	5—6
Plagioclase An 0-30	10—12	—	—
Epidote	2—3	—	—
Sericite	1—2	—	—
Apatite	1—2	—	5—6
Magnetic, other	2—3	—	6—7

a light brown-dark greenish brown dicroism the most developed crystals are in this rock-type. The greenish colourization appearing on the edges of the hornblende crystals can be regarded supposedly as an initial actinolitization (*Fig. 14*). Their perforated character is given by the biotite, apatite, sphene and pyroxene inclusons.

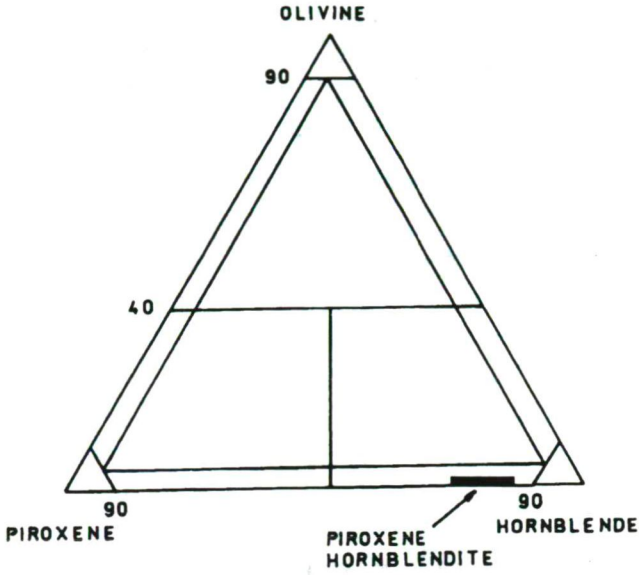

Fig. 13. Classification of ultramafic rocks containing hornblende: pyroxene hornblendite (after STRECKEISEN 1974).

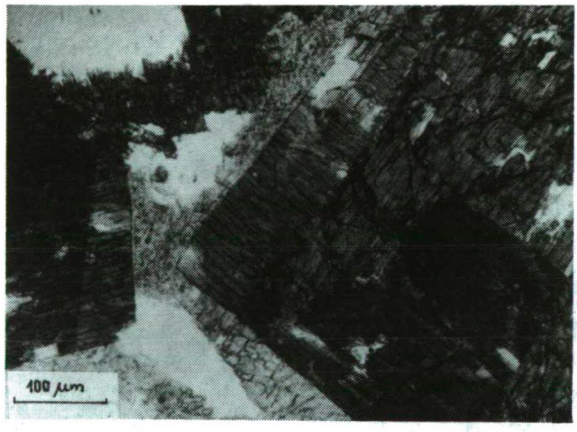

Fig. 14. Pyroxene hornblendite: slightly actinolitized hornblende. 28X, 1N.

The anhedrally shaped feldspars are observable as interstitial fillings among melanocratic minerals. The albite-oligoclase are normally non-twinned or occasionally exists a recognizable polysynthetic twinning. They contain a great number of euhedral and small prismatic pyroxene inclusions. The thick prismatic rock-forming pyroxenes are presented by augites (*Fig. 15*). The first generation sphenes form well-developed sometimes slightly weathered crystals. The biotite is also well-developed and second generation sphenes appear between the plates. The interstitial apatites show anhedral shape but there are euhedral apatite inclusions

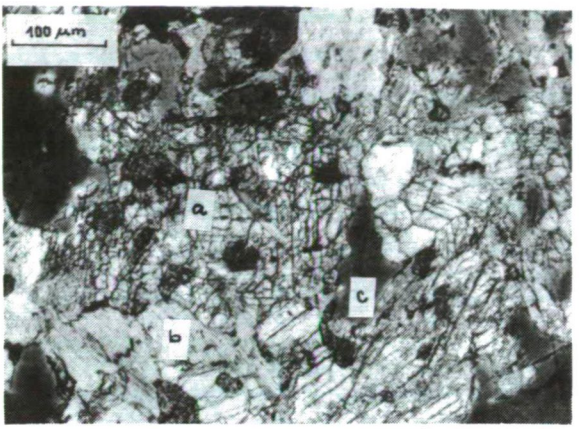

Fig. 15. Pyroxene hornblendite: (a) augite, (b) biotite, (c) chlorite. 28X, 1N.

in the melanocratic minerals. Magnetite occurs either in the form of large irregular grain penetrated by acicular apatite or as small euhedral crystal.

Pegmatoidic hornblendites are represented by smaller and bigger nests or lenses situated irregularly mostly in the microcrystalline hornblendites. Their extent can also reach 100—200 m. Petrographically they are composed by fairly well-developed hornblende and sphene crystals as well as biotite plates which are embedded into a slightly sericitized and non-twinned albite-oligoclase groundmass (CODARCEA *et al.* 1957). Locally small carbonatic patches are also recognized in them (*Fig. 16*). The biotite plates often contain fibrous rutile inclusions as a network originated by rutile twin-law (sagenite lattice).

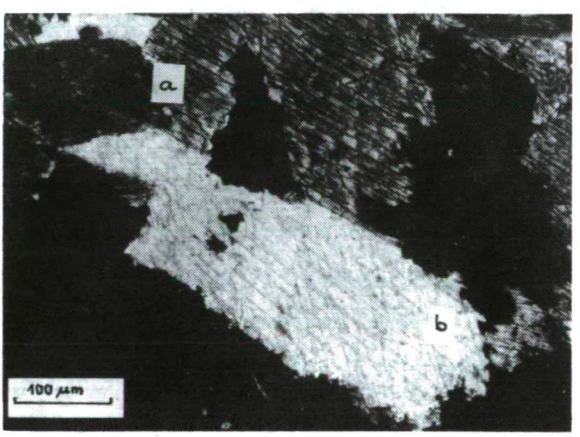

Fig. 16. Pegmatoidic hornblendite: (a) hornblende, (b) calcite. 28X, +N.

TABLE 5.

Modal composition of pegmatoidic hornblendites

Minerals	volume per cent
	8
Hornblende	30—75
Biotite	2—8
Plagioclase An 9-18	8—50
Apatite	10—12
Titanite	3—8
Carbonate	2—3
Sericite	1
Opaque	0—6

CONCLUSIONS

The first exact petrographic classification of the Northern Hornblendite Group of Ditró syenite massif is carried out in this paper. The groups of above-mentioned form not to sharply definable petrological units in space. Transitions or definite boundaries between the rock types equally occur. Genetic interpretation is missing yet. This reviewed mineralogical-petrographic description of Ditró (Orotva) hornblendites indicates magmatic-metasomatic processes in their development. However, a detailed genetical analysis requires an exact knowledge of spatial and temporal connections of these groups as well as determination of tectonic character and cycle of events in the massif. Most important tasks is to produce sufficient proofs for this analysis by systematically carried out investigations.

ACKNOWLEDGEMENT

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ON THE PYROXENE ANDESITES FROM THE TOKAJ MOUNTAINS AREA (NE HUNGARY)

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ABSTRACT

This paper deals with the geochemistry of the pyroxene andesite in selected three localities from the Tokaj Mountains, North East Hungary. Major elements for twenty two samples are reported and discussed to conclude the chemical characteristics of these andesites.

The distribution and behaviour of the trace elements in the representative examined andesites from the three localities were studied. The implications of the data for the origin of andesites are discussed.

INTRODUCTION

The Tokaj Mountains include two geologically independent units. One of them is the Vilyvitány Block. The other is the Szerencs Hill country (Inselberg).

The relatively large middle part of the Tokaj Mountains between the Bózsva and Szerencs brooks is still a genuine highland in the north, while the southern part is getting narrower and lower with a characteristically protruding outpost Mountain Kopasz at the town of Tokaj.

Among the members of the so-called Inner Carpathian volcanic belt in the territory of Hungary, the Tokaj Mountains have the most varied types of rocks. In addition to the "acidic" pyroxene anndesite, rhyolite and rhyolitic tuff (forming the main mass of the mountains), dacites and the so-called "true" pyroxene andesite cover large areas.

The majority of the pyroxene andesite can be found in two large and many small areas in the southeastern part of the mountains (*Fig. 1.*). In addition to the areas marked in *Fig. 1*, the rocks of Mulató Hill near Erdőbénye Village is signed as pyroxene andesite on the geological map of the Tokaj Mountains, but its rock has to named dacite (KULCSÁR et al. 1971; RÓZSA 1987).

The rocks of the three areas, Kopasz Hill near Tállya Village, Szokolya-Párkány Hill near Erdőbénye Village, Várhegy and its environs near Bodrogszegi Village and three samples from boreholes (Bsz.1, M.23, Eb.165) were studied.

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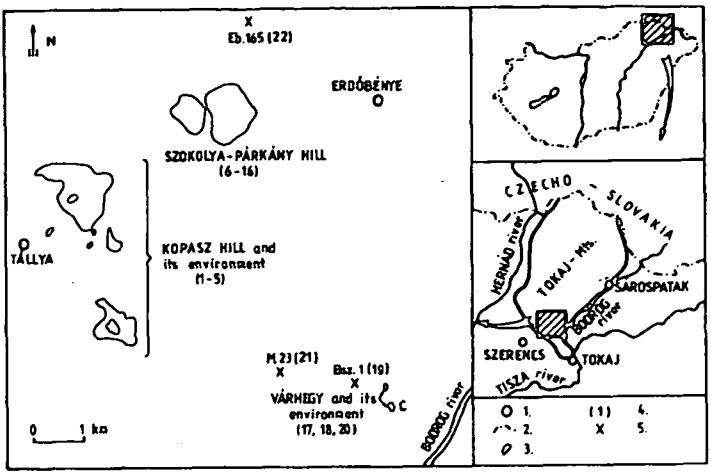

Fig. 1. Sketch map of the investigated area. 1) Settlement; 2) State boundary; 3) "True" pyroxene andesite; 4) Number of analysis and 5) Borehole.

PETROGRAPHY

Among the rocks of the three territories, the Szokolya-Párkány Hill near Erdőbénye Village is the most interesting one. This hill was formed by central volcanic eruptions (JUGOVICS 1965). The groundmass (<10 μm) form about 40 % of the rock, whereas the grains between 10—1000 μm and >1 mm represent 55 and 5 % respectively. The plagioclase feldspars are the most common phenocrysts with a composition ranging from An_{70} to An_{78}. The mafic minerals are represented by augite, less amount of hypersthene and olivine. The latter forms about 1-2 % of the whole rock with a composition ranging from Fo_{86} to Fa_{14} (PANTÓ,1970).

The rock of Kopasz Hill near Tállya Village has a relatively low quantity of the groundmass (approx. 20 %) due to its subvolcanic origin. The most abundant phenocrysts are plagioclase feldspars. The mafic minerals are represented by augite and hypersthene. Very rare small olivine grains could also be found. The texture of the rock is microholocrystalline porphyritic.

In the small cavities of the rock, several secondary minerals occur (e.g. calcite, dolomite, siderite, barite).

The andesite of Várhegy and its environs is also product of volcanic activity. The amount of the groundmass is about 40 % of the whole rock. The most abundant phenocrysts are plagioclase feldspars. The mafic minerals are represented also by both augite and hyperstene.

CHEMICAL CHARACTERISTICS

Major Elements Composition

The major element analyses of the samples from the investigated area are listed in Table 1. Table 2 shows these data recalculated on a H_2O and CO_2 free basis. In calculations, these values were used.

TABLE 1.

Major element analyses of the investigated rocks

	Kopasz hill near Tállya village					Szokolya-Párkány hill near Erdőbénye village											Várhegy and its environment near Bodrogszegi village				M. 23	Eb. 165
	1[a]	2[a]	3[a]	4[a]	5[b]	6[a]	7[a]	8[a]	9[a]	10[a]	11[a]	12[a]	13[a]	14[a]	15[a]	16[b]	17[c]	18[c]	19[c]	20[b]	21[a]	22[a]
SiO_2	57.06	57.53	56.71	58.36	55.87	55.50	54.56	55.65	54.34	54.86	56.58	52.80	53.79	53.58	53.74	55.40	53.07	54.74	54.10	53.29	54.67	57.06
TiO_2	1.20	1.18	1.22	0.95	1.16	0.63	0.38	0.81	0.47	1.02	0.40	0.45	0.42	1.06	1.18	0.80	1.27	1.21	1.50	1.25	0.90	0.73
Al_2O_3	18.52	17.31	18.39	15.82	17.46	18.96	18.23	17.00	17.33	16.31	18.79	19.21	18.57	17.23	16.76	18.12	18.80	18.39	16.50	17.84	19.34	15.57
Fe_2O_3	0.26	0.67	1.16	1.64	1.12	1.80	0.55	1.99	1.99	2.11	0.38	2.79	1.58	2.03	2.96	0.96	5.92	3.41	1.36	2.79	1.60	3.15
FeO	5.80	5.72	5.85	4.35	5.65	4.54	5.55	4.32	4.16	5.20	5.36	4.17	4.63	5.38	4.72	5.39	3.03	4.42	7.20	5.08	4.94	4.45
MnO	0.21	0.05	0.16	0.09	0.14	0.19	0.13	0.34	0.17	0.41	0.20	0.20	0.17	0.10	0.10	0.13	0.18	0.15	—	0.14	0.09	0.10
MgO	2.07	3.51	2.79	3.22	2.88	5.22	5.97	5.15	5.23	4.18	4.32	6.19	5.58	6.97	6.77	5.29	2.67	3.90	2.97	4.23	2.28	4.78
CaO	6.64	6.00	6.71	7.58	6.99	8.27	10.80	7.53	8.26	9.89	8.85	8.95	8.90	9.77	9.78	8.73	7.91	8.10	8.04	9.09	8.43	8.14
Na_2O	3.46	3.30	2.55	2.38	3.17	2.36	2.18	2.44	2.34	2.56	2.40	2.41	2.04	2.36	2.39	2.73	2.86	3.05	3.14	3.30	3.18	1.99
K_2O	1.93	2.18	2.04	0.68	2.50	1.45	1.24	1.85	1.78	1.32	1.49	1.32	1.35	1.20	1.12	1.97	1.33	1.37	1.12	1.56	1.60	1.43
P_2O_5	0.23	0.25	0.26	0.13	0.29	0.15	0.10	0.71	0.13	0.10	0.13	0.15	0.18	0.25	0.25	0.20	0.23	0.22	—	0.23	0.24	0.12
CO_2	1.34	0.62	1.00	1.98		—	—	—	0.19	0.08	—	0.06	—	0.17	0.21		—	—	—		1.14	0.04
$+H_2O$	1.35	1.92	1.83	2.23	2.16	1.18	0.70	1.27	2.47	1.24	0.81	1.40	2.35	0.42	0.35	0.65	2.22	1.13	1.15	0.96	0.88	1.20
$-H_2O$	0.24	0.12	0.12	0.25		0.30	0.17	1.06	1.61	0.37	0.25	0.45	0.76	0.27	0.23		0.61	0.41	3.40		0.72	1.04
Σ	100.31	100.36	100.79	99.66	99.39	100.55	100.56	100.12	100.47	99.65	99.96	100.55	100.32	100.79	100.56	100.37	100.10	100.50	100.48	99.76	100.01	99.80

[a] data from Gyarmati (1977)

[b] new analysis; collector: Rózsa P., analyst: Nazih Aly Saad

[c] data from Rózsa-Barta (1986)

TABLE 2.

Major element contents of the investigated rocks recalculated on a H_2O and CO_2 free basis

	Kopasz hill near Tállya village					Szokolya-Párkány hill near Erdőbénye village											Várhegy and its environment near Bodrogszegi village				M. 23	Eb. 165
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22
SiO_2	58.60	58.88	57.96	61.30	57.46	56.02	54.73	56.91	56.49	56.00	57.21	53.53	55.33	53.62	53.86	55.56	54.56	55.32	56.40	53.94	56.21	58.51
TiO_2	1.23	1.21	1.25	1.00	1.19	0.64	0.38	0.83	0.49	1.04	0.40	0.46	0.43	1.06	1.18	0.80	1.31	1.22	1.56	1.27	0.93	0.75
Al_2O_3	19.02	17.72	18.80	16.62	17.96	19.14	18.29	17.38	18.01	16.65	19.00	19.47	19.10	17.24	16.80	18.17	19.33	18.58	17.20	18.06	19.88	15.97
Fe_2O_3	0.27	0.69	1.18	1.72	1.15	1.82	0.55	2.03	2.07	2.15	0.38	2.83	1.63	2.03	2.97	0.96	6.09	3.45	1.42	2.82	1.64	3.23
FeO	5.96	5.85	5.98	4.57	5.81	4.58	5.57	4.42	4.32	5.31	5.42	4.23	4.76	5.38	4.73	5.41	3.11	4.47	7.50	5.14	5.08	4.56
MnO	0.21	0.05	0.16	0.10	0.15	0.19	0.13	0.35	0.18	0.42	0.20	0.20	0.17	0.10	0.10	0.13	0.18	0.15	—	0.14	0.09	0.10
MgO	2.12	3.59	2.85	3.38	2.96	5.27	5.99	5.27	5.44	4.27	4.37	6.28	5.74	6.98	6.79	5.30	2.74	3.94	3.10	4.28	2.34	4.90
CaO	6.82	6.14	6.86	7.96	7.19	8.35	10.83	7.70	8.59	10.10	8.95	9.07	9.16	9.78	9.80	8.75	8.13	8.19	8.38	9.29	8.67	8.35
Na_2O	3.55	3.38	2.61	2.50	3.26	2.38	2.19	2.49	2.43	2.61	2.43	2.44	2.10	2.36	2.40	2.74	2.94	3.08	3.27	3.34	3.27	2.04
K_2O	1.98	2.23	2.08	0.71	2.57	1.46	1.24	1.89	1.85	1.35	1.51	1.34	1.39	1.20	1.12	1.98	1.37	1.38	1.17	1.58	1.64	1.47
P_2O_5	0.24	0.26	0.27	0.14	0.30	0.15	0.10	0.73	0.13	0.10	0.13	0.15	0.19	0.25	0.25	0.20	0.24	0.22	—	0.23	0.25	0.12
Σ	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00

The data show that the andesite of Kopasz Hill has the highest SiO_2 content, followed by the andesite of Szokolya-Párkány Hill and Várhegy and its environs respectively. The highest TiO_2 and iron contents are in the andesite of Várhegy and its environs. Corresponding mineral composition, the andesite of Szokolya-Párkány Hill has the highest MgO. In accordance with the SiO_2 content, rocks of Szokolya-Párkány Hill and Várhegy and its environs have a higher CaO content than that of Kopasz Hill, whereas alkalis have a reversed distribution.

The foregoing mentioned trends are clearly shown in the FMA diagrams (*Fig. 2*); the points of the three areas clearly separated from each other, only one point of Szokolya-Párkány Hill and another one of Kopasz Hill fall into different field (iron-rich field). In the CNK diagram (*Fig. 3*) grouping of the points by the three areas is obvious.These tendencies are shown together by the F-M diagram (*Fig. 4*) of SIMPSON (1954). The andesite of Szokolya-Párkány Hill has the lowest M and F values because of its high MgO and CaO contents, whereas the andesite

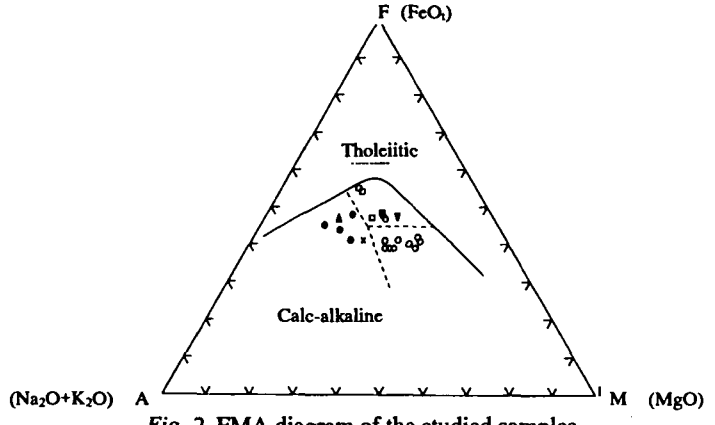

Fig. 2. FMA diagram of the studied samples.
Symbols: ● Kopasz Hill, ○ Szokolya-Párkány Hill, □ Várhegy and its environment, ▲ M. 23 borehole, ▼ Eb. 165 borehole and X average of the acidic andesite.
FeO_t = total iron as FeO.

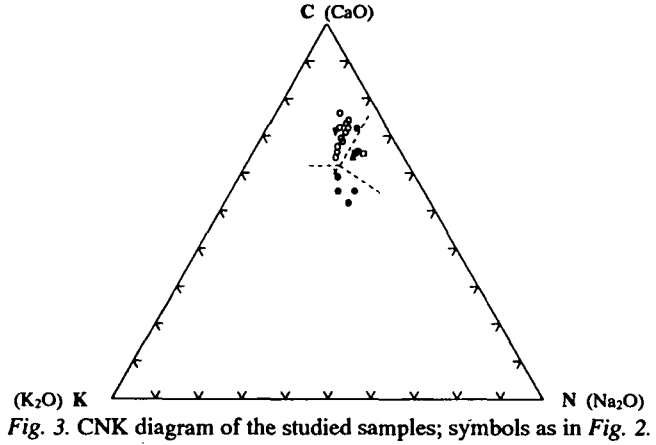

Fig. 3. CNK diagram of the studied samples; symbols as in *Fig. 2.*

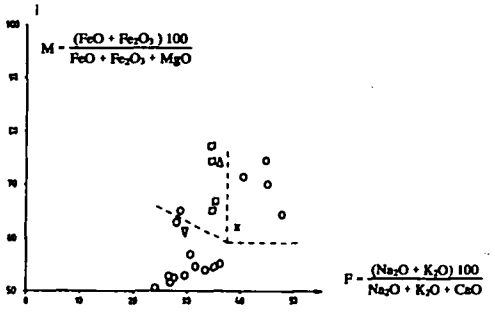

Fig. 4. F-M diagram of the studied samples; symbols as in *Fig. 2.*

of Kopasz Hill has highest M and F values because of the relatively low quantities of MgO and CaO. In the case of the points of Várhegy and its environs, M values are similar to that of Kopasz Hill and F values are similar to that of Szokolya-Párkány Hill. The numerical values of the MgO/FeO_t ratios of Kopasz Hill and Várhegy and its environs are similar to each other and lower then that of Szokolya-Párkány Hill (*Fig. 5*). However, the points of the three areas are well distinct.

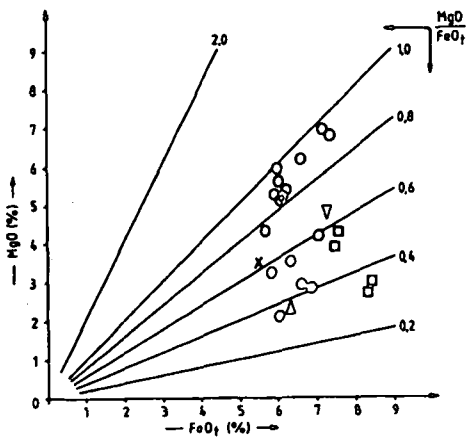

Fig. 5. MgO vs. FeO_t diagram of the studied samples; symbols as in *Fig. 2.*
FeO_t = total iron as FeO.

On the basis of the CaO/Alk ratios (*Fig. 6*), the points of Kopasz Hill are well separated from the other two groups. On the other hand, and on the basis of the Na_2O/K_2O ratios the separation of the three groups can be distinctly seen (*Fig. 7*).

It is noteworthy that the average point of the so-called "acidic" andesite is close to the points of Kopasz Hill in each diagram.

Classification and Nomenclature

Recently, there are two nomenclature diagrams accepted by IUGS for the classification of igneous rocks. The QAPF diagram (STRECKEISEN 1980) and the so-called TAS diagram (LE BAS *et al.* 1986) which distinguishes the volcanic rocks on the basis of their silica and alkalis contents. In our case, the latter was applicable because the groundmass of the rocks can reach about 40 % and these rocks do not contain either quartz or feldspathoids, therefore the location of all point could be found on the AP line on the QAPF diagram (diagram not shown here). It can be

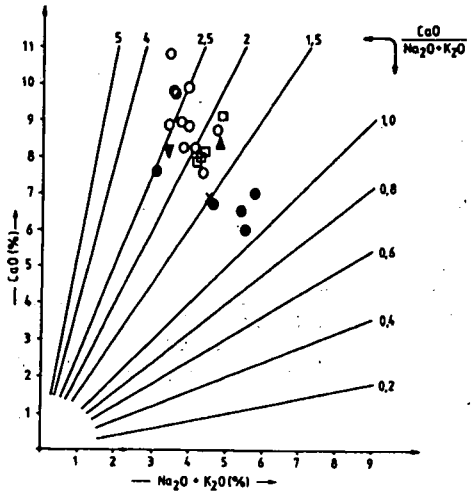

Fig. 6. CaO vs Alk. diagram of the studied samples; symbols as in *Fig. 2.*

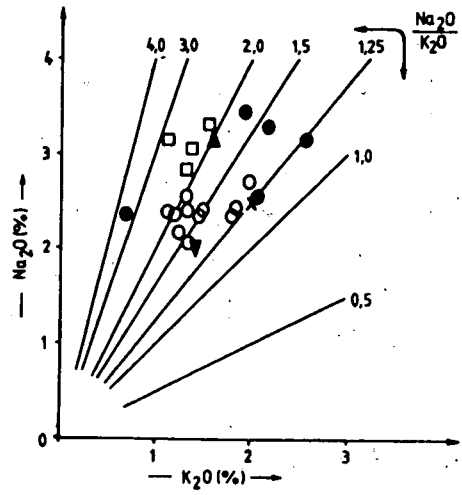

Fig. 7. Na_2O vs K_2O diagram of the studied samples; symbols as in *Fig. 2.*

seen from Figure 8 that each sample of Várhegy and Szokolya-Párkány Hill is classified as basaltic andesite, except sample 11, while that of Kopasz Hill fall into the andesitic field. These results are confirmed by the classification of TAYLOR *et al.* (1981) as shown in *Fig. 9*, where the same results have been obtained.

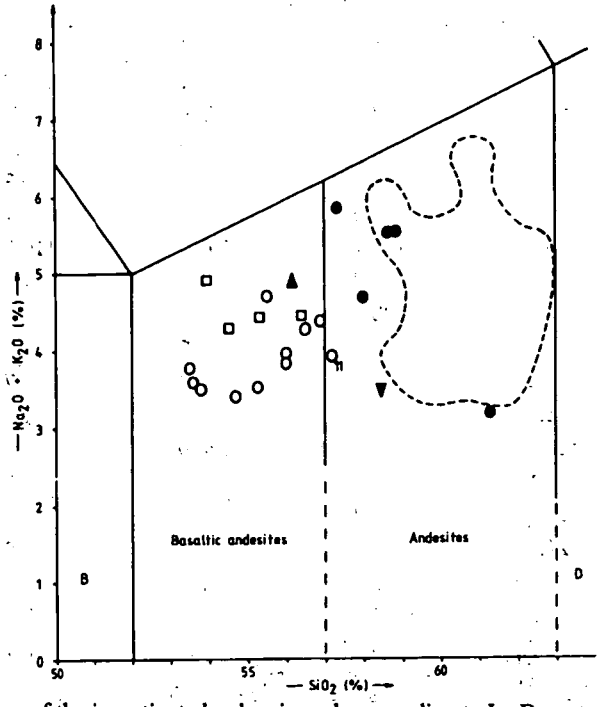

Fig. 8. Classification of the investigated volcanic rocks according to LE BAS *et al.* (1986); symbols as in *Fig. 2.*

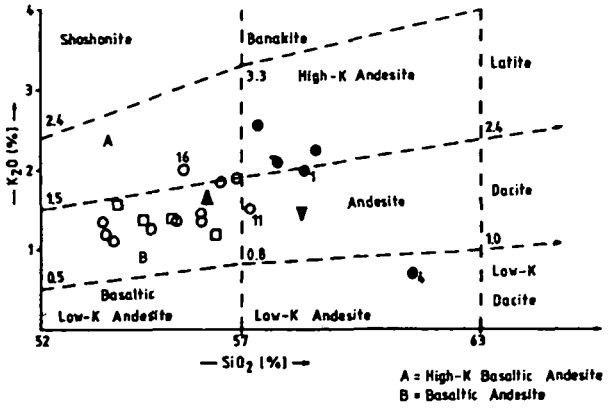

Fig. 9. Classification of the investigated volcanic rocks according to their K_2O content as suggested by TAYLOR *et al.* (1981); symbols as in *Fig. 2.*

Trace Elements Chemistry

15 trace elements were determined in the samples in question by ARL-34000-ISC emission spectrometer equipment (Utrecht, Holland).

The trace element analyses of representative samples from the three hills together with the average trace elements in andesite of TAYLOR *et al.* (1969) are listed in Table 3.

TABLE 3.

Trace element concentrations of the representative samples of the investigated andesitic rocks

	1	2	3	*
Sr	379.8	314.6	324.1	385.0
Ba	463.0	378.2	252.6	270.0
Ce	42.5	46.6	40.7	24.0
Sn	<2.6	<2.5	<2.7	0.8
Mo	<0.7	<0.6	<0.7	1.1
Zr	143.8	183.7	150.7	110.0
Li	17.4	9.7	14.8	—
Cu	21.1	20.2	15.2	54.0
Zn	71.6	78.2	76.2	—
Pb	<10.0	<9.7	<10.4	6.7
Co	18.3	16.3	22.9	24.0
V	148.0	147.8	176.9	175.0
Ni	18.1	10.4	20.6	18.0
Cr	58.3	16.2	68.1	56.0
Y	24.9	32.9	30.9	21.0
Ba/Sr	1.20	1.20	0.80	0.70
V/Ni	8.20	14.20	8.60	9.70
Ni/Co	0.99	0.64	0.90	0.75

1. Erdőbénye, Szokolya-Párkány Hill. 2. Tállya, Kopasz Hill, Quarry.
3. Bodrogszegi, Várhegy. * Average andesite of TAYLOR *et al.* (1969).
all concentration in ppm.

The compatible trace elements Co, Ni, Cr which are generally incorporated in ferromagnesian minerals and generally decrease with increasing differentation, show higher concentration values in the pyroxene andesite of Várhegy and decrease toward the Kopasz andesite through the andesite of Szokolya-Párkány Hill, and this is true if we compare these concentration with their major elements content in the foregoing FMA diagram and these values (Table 2) are in agreement with the average values of TAYLOR *et al.* (1969), though V, Cu, Zn and Y content show narrow range of variation in the three hills.

Zr content, on the other hand, varies highly and shows the highest concentration value in the andesite of Kopasz Hill (183.7 ppm) and this value is higher than the average value (110 ppm) given by TAYLOR *et al.* (op. cit).

The distribution of Ba and Sr in the examined andesite indicates that Ba is more abundant than Sr in Szokolya-Párkány and Kopasz Hill. The highest concentration of Ba and Sr is recorded in Szokolya-Párkány andesite and represented by 436 and 379 ppm respectively due to the high value of K_2O and CaO.

The examined andesite is characterized by low abundance of cobalt and nickel, and high abundance of vanadium. Ni/Co ratios for the three localities are less than one, and V/Ni ratios show the reverse relationship as being greater than 8.

The low values of Ni and Cr of the three localities perclude the derivation of andesites by mixing between basaltic magma and acidic material. According to TAYLOR *et al.* (1969), the high V/Ni ratios perclude the derivation of calk-alkaline andesites from alkali or tholeiitic basalts by crystal fractionation at low pressure. Since under these conditions, vanadium will be depleted in the residual magma as well as nickel. The high vanadium contents also perclude the operation of the mechanism proposed by OSBORN (1962) to account for the lack of iron enrichment in calk-alkaline rocks. On the other hand, the Ni/Co ratios of examined andesite increase with increasing K_2O or SiO_2. The decrease of Ni/Co ratios with the increasing K_2O or SiO_2 is considered as an evidence that andesitic rocks were derived by fractional crystallization from basic magma (TAYLOR *et al.* 1969).

Accordingly, it can be concluded from the behaviour of the trace elements and their ratios in the three localities (low contents of Ni, Cr, low Ni/Co ratios (<1) and high V contents) that the examined andesites represent derivatives of primary andesitic magma.

MAGMA TYPES AND TECTONIC IMPLICATION

The AFM diagram (IRVINE and BARAGAR 1971) has widely used to differentiate between tholeiitic and calk-alkaline magmas (*Fig. 2*). From this figure the examined andesites display a clear calk-alkaline affinity.

PEARCE *et al.* (1977) have proposed a FeO_t, MgO, Al_2O_3 triangular diagram to relate the chemical composition of basic and intermediate rocks with their tectonic environment. These authors screened out the alkaline rocks in order to improve their relationship with the tectonic environment. *Fig. 10* shows such a plot for the analysed samples. A clear tendency for the orogenic and spreading center island environments, can be seen from the plot.

According to the diagram after PEARCE (1980) and as shown in *Fig. 11*, the examined rocks located in the island arc basalt field.

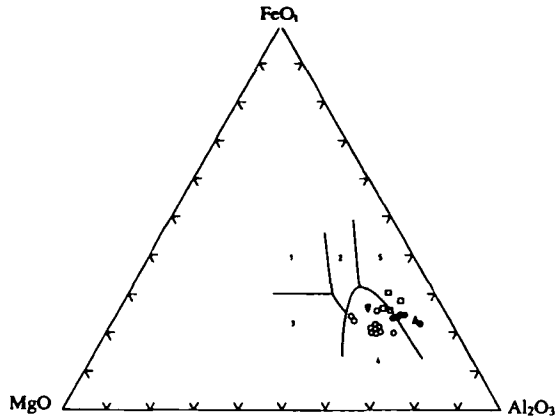

Fig. 10. Rock analyses plotted on the FeO_t-MgO.Al_2O_3 diagram. The fields numbered 1-5 are: 1) ocean island 2) continental 3) ocean ridge and floor, 4) and 5) orogenic and spreading center island respectively; symbols as in *Fig. 2*.FeO_t = total iron as FeO

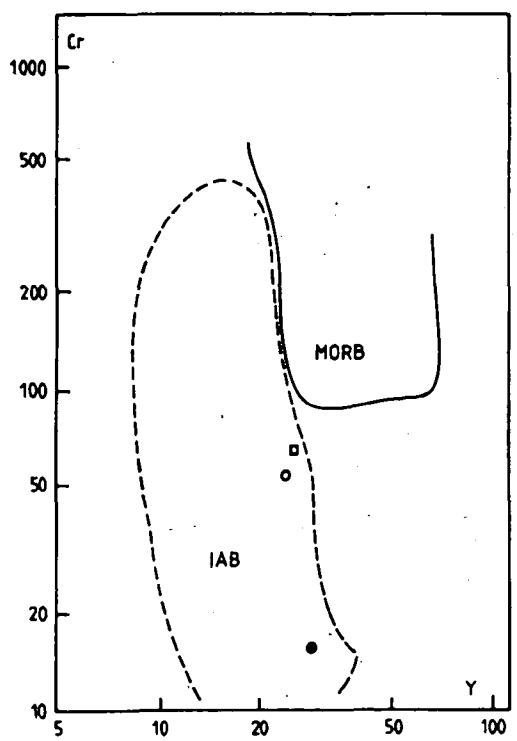

Fig. 11. Cr vs Y diagram (after PEARCE 1980) MORB = solid ballon, and IAB = dashed ballon, indicate fields of mid-ocean ridge basalts and island-arc basalts respectively; symbols as in *Fig. 2.*

CONCLUSION

On the basis of the studies on the so-called "true" pyroxene andesite from southern part of Tokaj Mountains, the following can be concluded:

(1) On the basis of the recommendation of TAS diagram accepted by IUGS, the name "basaltic andesite" should be used instead of the name "true" andesite. Rocks of Szokolya-Párkány Hill and Várhegy and its environs have to be arranged into this group. "True" andesite of Kopasz Hill, however, together with "acidic" andesite, has to named simply as andesite.

(2) Classificational differences are incident to the chemical compositional differences of the rocks. It is characteristic that basaltic andesite of Szokolya-Párkány Hill and Várhegy and its environs have lower total alkalis compared to the iron and MgO content and have higher CaO content than andesite of Kopasz Hill. Because of their CaO content, the basaltic andesite samples have relatively low F indices. The close connection between "acidic" andesite and the andesite of the Kopasz Hill is proved by the fact that the average point of "acidic" andesite is close to the points of andesite of Kopasz Hill in each diagram.

(3) Although both rocks of Szokolya-Párkány Hill and that of Várhegy and its environs should be named basaltic andesite, there is a definite difference between their major elements composition besides the well known difference between their mineral assemblages. The olivine-bearing basaltic andesite of Szokolya-Párkány Hill has slightly lower Na_2O and iron content than the basaltic andesite of Várhegy and its environs. In accordance with this fact, the basaltic andesite of Várhegy and its environs have lower MgO/FeO_t ratio and M index. Consequently, it seem to reasonable to subdivide the basaltic andesite of the Tokaj Mountains into two groups: (1) the less differentiated "Szokolya"-type and (2) the more differentiated "Várhegy"-type. The petrochemical differences of the two groups are as follows:

TABLE 4

	FeO_t	MgO	Na_2O	MgO/FeO_t	M-index
Szokolya-type:	<7	>4	<2.7	>0.6	<68
Várhegy-type:	>7	<4	>2.7	<0.6	>68

Sample of Eb. 165 borehole has to be named andesite, rock of the M. 23 borehole is basaltic andesite and belongs to "Várhegy"-type.

From the trace elements and their ratios, it is obvious that the investigated andesites represent derivatives of primary andesitic magma.

It was found that the three areas under consideration have mainly calk-alkaline affinity, indicating an island arc tectonic setting.

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